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GEOLOGY

PRINCIPLES AND PROCESSES

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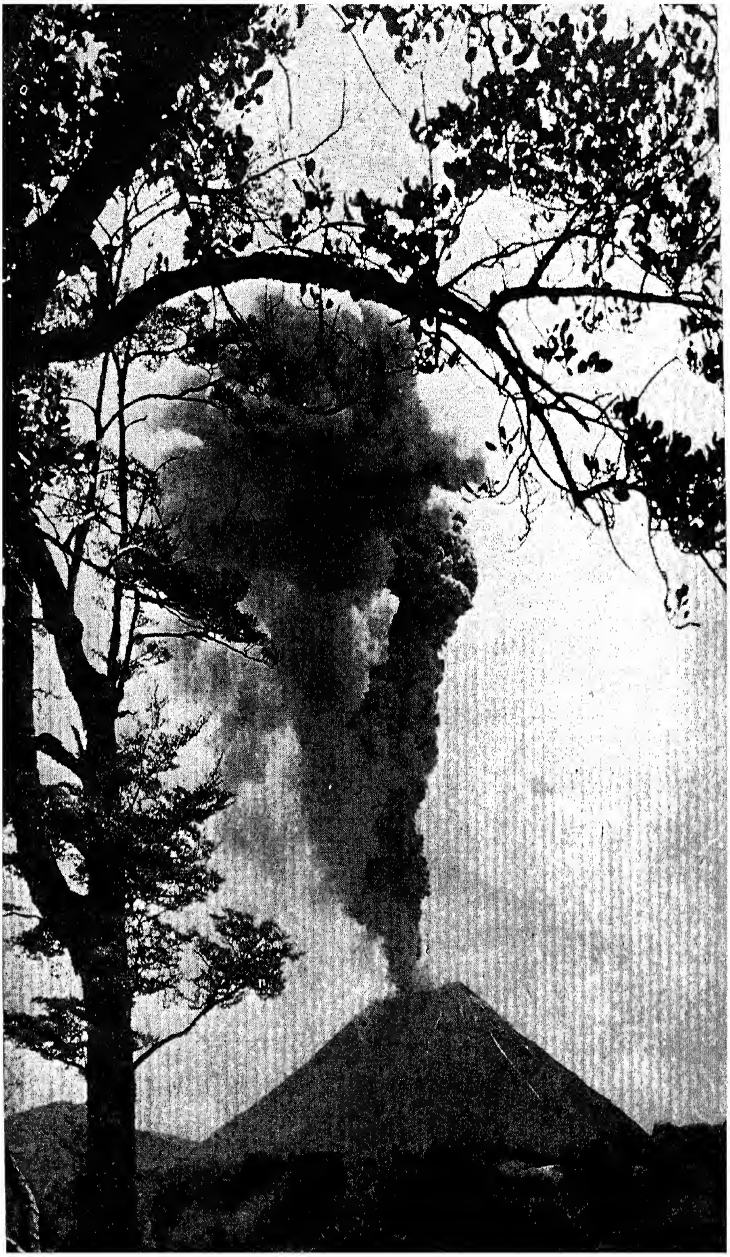

Frontispiece.—The Volcano Ngauruhoe, New Zealand. (*Copyright National Geographic Society. Reproduced with permission.*)

GEOLOGY

PRINCIPLES AND PROCESSES

by

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PREFACE TO THE SECOND EDITION

In its second edition, "Geology: Principles and Processes" (originally published under the title "Geology") has been revised and completely reset. The chapter on the atmosphere and the work of the wind has been reorganized and enlarged, particularly the section treating movements of the atmosphere. This seemed to be desirable because of the interest of the public in aviation. The chapters on weathering and on the atmosphere have been transposed. This brings the section on weathering and on ground water in sequence; it is believed that this arrangement is sound because the two subjects are so closely related. The discussion of the geysers has been transferred from the chapter on vulcanism to that on ground water. The treatment of varved glacial lake sediments and their importance in glacial and postglacial chronology has been expanded. A chapter on sediments and realms of sedimentation has been added, and the chapter on sedimentary rocks has been reorganized and expanded. The chapter on vulcanism has been divided into two chapters, one on vulcanism and another on igneous rocks.

The chapter on Diastrophism has been enlarged to include Structures of Rocks, which in the first edition was treated in a separate chapter. This makes a long chapter, since little if any material has been taken out, and some material has been added on both subjects. It seemed logical, however, to bring the various diastrophic processes and their results together in a single chapter. The discussion of earthquakes has been abridged.

The chapter on Probable Conditions within the Earth and that on Mineral Deposits have been transposed, an arrangement that is believed to be an improvement pedagogically.

The chapter on earth history (formerly Chapter XVIII) has been omitted. Space was not available for an adequate treatment of the subject, and, judging by the arrangement of most geology courses, it seemed out of place.

Essentially all chapters have been expanded. The second edition is about as large as the first edition, even with the omission of the chapter on earth history. The authors believe that the reader will find the illustrations an improvement over those of the first edition in respect to both choice of subject and reproduction.

THE AUTHORS.

MINNEAPOLIS, MINN.,
December, 1938.

PREFACE TO THE FIRST EDITION

The textbook "Geology" is a brief presentation of the subject, prepared for use in a one-semester beginning course for college students. It is intended to give the student some knowledge of the materials of the earth and of the processes that operate at the earth's surface and that have operated in the past to form the earth.

The analysis of land forms and the influence of subsurface structures on surface features cannot be fully appreciated by students of geology until they have some conception of the nature of the materials of which the earth is composed. For this reason the introductory chapter is followed by a very brief treatment of common minerals and rocks. For a fairly comprehensive course in geology it is desirable that a certain amount of laboratory work support the course, or, if that is impracticable, specimens of common minerals and rocks should be available for study during the early part of the course.

The early chapters treat the processes that in a large measure come under observation, whereas the later ones treat the processes that are less commonly observed. The current theories regarding the earth's interior are more difficult for the student to follow and in the main these are grouped in the chapter on Probable Conditions within the Earth. In the briefer courses this chapter may be omitted. In certain curricula geology is presented in two consecutive courses such as (1) geologic processes and (2) historical geology. With such a division it would probably be desirable to omit the last chapter of the text, which is a brief outline of historical geology. In curricula where geology is treated in the two consecutive courses, (1) geologic processes and (2) elementary economic geology, it would be desirable to include the chapter on historical geology.

In general a beginning course in geology is given to a group of students with various kinds of preparation for the work. As a rule, a few of the class will have had some training in chemistry, but many will have had none. It is very difficult to treat certain geological processes without the use of elementary chemistry and to describe minerals, rocks, and rock alteration without utilizing formulae. This is the most difficult pedagogical problem that confronts the instructor of geology or the writer of a book on the subject. We believe the problem is best solved by some use of chemistry in discussion of geologic processes when that is necessary and by the presentation of formulae and some of the simple equations in footnote references. It is impracticable to require the student who has had no instruction in chemistry to learn these equations, yet they should be readily available to the student who has some knowledge of chemistry. Certain instructors give a few lectures on chemistry in an elementary course in geology; when this is done it is convenient to have equations readily available.

Most students of an elementary course do not consistently use the dictionary, the atlas, and the encyclopedia. It is desirable, therefore, to introduce definitions, index maps, and drawings where these will simplify study, and this has been done as far as practicable.

In the preparation of the text all writers collaborated on Chapter I (Introduction) and on Chapter II (Materials of the Lithosphere). The chapters on atmosphere and on weathering are by Allison. The chapters on underground water, on running water, on glaciers, on lakes and lake deposits, and on mountains are by Thiel. The chapters on the ocean, on sedimentary rocks, and on the outline of historical geology are by Stauffer. The chapters on diastrophism, on vulcanism, on metamorphism, on structural geology, on ore deposits, and on the interior of the earth are by Emmons.

We have endeavored to acknowledge sources of information by text and footnote references and by references placed at the end of chapters. In an elementary text it is impracticable, however, fully to acknowledge all important sources of information, although that is desired. We have aimed to acknowledge scrupulously the sources of illustrations and of drawings used or from which we have obtained ideas that are incorporated in new drawings.

In connection with the preparation of drawings, we have found helpful the works of A. K. Lobeck and of Barrington Brown and Debenham. The drawings were prepared chiefly by Mr. H. R. Norman, Mr. W. I. Gardner, Mr. L. J. Snell, Mr. A. C. Papas, Miss Jane Titcomb, and Miss Faith Patterson, to whom we are grateful for their excellent workmanship.

We wish especially to express our indebtedness to our colleagues, Doctors F. F. Grout, J. W. Gruner, G. M. Schwartz, M. H. Frohberg, and R. B. Ellestad for many valuable suggestions and for other courtesies in connection with the preparation of the text.

THE AUTHORS.

MINNEAPOLIS, MINN.
September, 1932.

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CHAPTER I

INTRODUCTION

Geology is the science which treats of the history of the earth and its inhabitants. It is the interpretation of the sequence of events from the beginning of the earth as a definite planet in the solar system, through its many changes during the long geologic ages, to the present time. It is concerned with the composition, character, and architecture of the earth's crust and with the agencies and processes which continually are altering it. Geology enlists the aid of other sciences and attempts to interpret the earth in the light of our knowledge of chemistry, physics, astronomy, biology, and other sciences.

Geology includes the study of rocks and their relations to each other; the study of water and its effect upon the rocks; the study of the atmosphere, its movements, and its reactions with the mineral constituents of rocks. A geologist is called upon not only to classify minerals and rocks but to locate valuable mineral deposits and to direct their exploration and to give counsel regarding dam sites, tunnels, foundations, irrigation systems, control of floods, and other engineering projects.

The earth is continually changing. Some changes are great and rapid, whereas others are small and slow although they are nevertheless effective. The geologist seeks to analyze the changes and the processes that cause them. He assumes that these processes have been operative during the past as well as in the present, although perhaps in different degrees, and he seeks to explain the present earth as the result of processes which have been acting through long ages of time. Thus the present, which is the outgrowth of the past, is also the key to the history of the past.

Some geologists devote themselves largely to the study of minerals and rocks; others to the study of the arrangement of rocks, of fossils, or of ore deposits; and still others to the study of the history of the earth. For this reason the science of geology commonly is divided into several branches, each of which emphasizes certain phases of the subject. *Cos-*

mology treats of the early history of the earth and the relations of the earth to other heavenly bodies in the universe, such as the sun, the other planets, and other stars. *Petrology* treats of the rocks of the earth. *Structural geology* deals with the arrangements or the structural relations of rocks and particularly with their relations to each other. *Dynamic geology* treats of the forces and the movements that have affected the rocks and the results of these movements. *Physiography* deals with the development of the form or contour of the surface of the earth and the origin of the mountains, valleys, and plains. *Paleontology* is the study of the remains of ancient life that are found in the rocks. *Historical geology* is the study of the history of the earth as shown by its rocks and particularly the record of events that is revealed in the rocks and by the relations of the rock formations to each other. *Economic geology* treats of the occurrence, origin, and distribution of the materials of the earth that are valuable to man. It includes the study of deposits of the metals, coal, petroleum, and many other substances. All of these branches of geology are closely related to each other and form parts of the general subject of geology.

The main contribution of geology to human affairs has been in discovering and making available many of the raw materials utilized in industry. The extent of the uses of these materials is indicated by the fact that mining and allied activities represent one of the four basic industries which furnish the raw material needed in our modern world, the other basic industries being agriculture, lumbering, and fishing. According to the U. S. Bureau of Mines, the total value of the mineral production of the United States in 1937 was $5,440,000,000. Of this amount $1,444,400,000 was derived from metals (iron, copper, lead, zinc, gold, silver, aluminum, manganese, platinum, etc.), $3,122,900,000 from the mineral fuels (coal, natural gas, and petroleum), and $841,700,000 from miscellaneous non-metallic materials such as clay products, cement, building stone, crushed rock, sand and gravel, gypsum, lime, sulphur, salt, and slate. An abundance of mineral products is necessary to our complex industrial civilization, and a lack of them is a handicap to any people. The control of essential minerals profoundly affects the welfare of nations both in peace and in war.

Place of the Earth in the Universe.—When science was in its infancy, the earth was believed to be the center of the universe and to have the sun, moon, and stars moving around it in the arched heavens. Astronomy, however, shows that the earth is but a tiny speck of matter in the vastness of space. It plays a very inconspicuous part in the universe. Millions of larger masses called stars also are moving about in this space, and some are at such great distances that their movements are not readily perceived. These are known as fixed stars. Close observations indicate that these also are rushing through space about some common center. The Milky Way is a broad zone in which the stars of our galaxy seem to be concentrated.

One star is so near the earth that it appears like a great flaming globe. That star is the sun. There are thousands of stars similar to the sun and thousands of millions of others that differ from it as to size, mass, temperature, and age. Many stars are bigger and brighter than the sun. The average star has a volume a million times that of the earth, and some stars are a million million times larger (Fig. 1). Measured by the scale of the distances between the stars, the sun is near the earth. Its mean distance is 93,000,000 miles away. Such a distance taxes the imagination. If a bullet fired from a gun at the sun kept its muzzle velocity unimpaired, it would take 7 years to reach the earth. The distances between stars are far greater and are measured by a unit of length called a light year. One light year is the distance light travels in 365¼ days, traveling at a velocity of 186,000 miles per second.[1] The average

Fig. 1.—The great nebula in Andromeda. An island system smaller in size but similar in shape to our galactic system. The great nebula is over 50,000 light years in diameter and contains millions of stars many of which are thousands of times brighter than the sun. (*Photograph Courtesy Yerkes Observatory.*)

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[1] It takes the light from the sun only 8 minutes 19 seconds to reach the earth.

distance between stars is from 8 to 10 light years. In order more readily to visualize the enormous distances in the stellar system, consider the group of stars known as the Great Cluster of Hercules, which is 36,000 light years distant. The light which we receive from that cluster today left its surface 360 centuries ago. If a news dispatch could be flashed by radio from that star to the earth today, it would be 360 centuries old before it reached the earth.

Solar System.—The solar system (Fig. 2) consists of one star—the sun—about which revolve nine smaller spheres—the planets—of which the earth is one, a thousand or more smaller bodies known as asteroids or planetoids that occupy definite positions mainly between the planets Mars and Jupiter, and 27 satellites or moons that accompany certain planets. Some planets have no moon; the earth has one, Jupiter 9, and Saturn 9. All of the planets together do not equal one-fifth of 1 per cent of the mass of the sun. Thus nearly all of the mass of the solar system is in the sun.

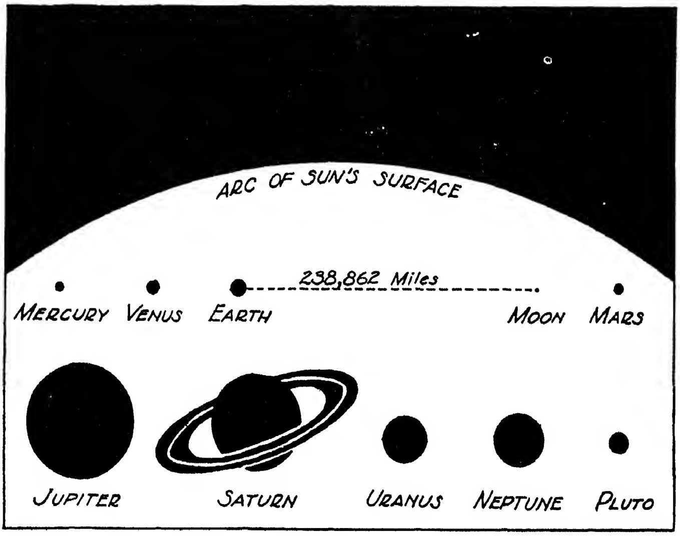

FIG. 2.—Diagram showing the major bodies of the solar system. Scale is shown by distance from earth to moon. (*Chiefly from Moulton.*)

All of the planets revolve approximately in the same plane and in the same direction about the sun (Fig. 3). Their average rate of speed around the sun is approximately 13 miles per second. Since the outer planets have less speed and greater distances to travel, their periods of revolution around the sun are longer. The sun also is rushing through space in the general direction of the bright star Vega at the rate of 12 miles per second. Therefore the planets and their satellites are carried with the sun as they continue their journey around it. Along with their orbital movements the planets rotate on their polar axes. The earth's period of rotation is 24 hours. This period of rotation causes its surface at the equator to move at the speed of about 17 miles a minute.

The earth's satellite is the moon. The diameter of the moon is about 2,160 miles, and the earth's mass is about eighty-one times that of the moon. The moon is about 240,000 miles away from the earth, which is less than ten times the distance around the earth. It is so near that its pitted surface is easily visible with a small telescope. These pits are supposed to be craters formed either by the infall of meteorites or by volcanism. Because the moon is so near the earth its gravitative pull upon the earth is greater than that of other heavenly bodies except the sun. Its pull upon the side of the earth that is nearest it is greater than its pull on the far side, and the tides are a result of this difference.

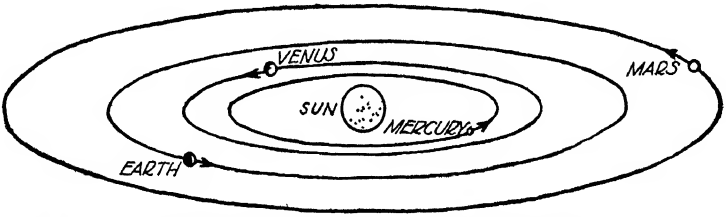

Fig. 3.—Diagram showing the four minor planets revolving about the sun. All of the nine planets revolve in approximately the same plane. Their orbits are elliptical.

All of the planets are far away from the earth, and little is known of their surfaces. A powerful telescope, however, shows a dark equatorial zone around Mars and a bright spot at its pole which possibly represents an ice cap. The satellites, or "moons," of the planets and also the rings of Saturn may be seen through a telescope.

Facts Concerning the Solar System
(*From Luyten, Moulton, and others*)

Sun and planets	Mean distance from the sun		Period of revolution, days	Diameter, miles	Density, water = 1
	Millions of miles	Earth = 1			
Sun	.....	.....	......	865,000	1.41
Moon	.....	.....	......	2,163	3.34
Mercury	36	0 387	88	3,030	3.80
Venus	67	0.723	225	7,700	4.85
Earth	93	1.000	365	7,918	5.52
Mars	142	1.524	687	4,230	4.01
Jupiter	483	5.20	4,333	86,500	1.33
Saturn	886	9.54	10,759	70,000	0 73
Uranus	1,782	19 19	30,686	31,500	1.22
Neptune	2,792	30 07	60,188	34,800	1.41
Pluto	3,673	39.50	92,611	4,000	

Major Divisions of the Earth.—The earth is regarded as consisting of three parts: (1) the atmosphere, or gaseous envelope, which surrounds the globe; (2) the hydrosphere, or water envelope, which surrounds the larger part of the solid globe and penetrates the rock sphere; (3) the lithosphere, or rock sphere, which constitutes the solid earth.

These three divisions are not independent; the atmosphere contains water and solid rock particles, or dust; the hydrosphere absorbs air and carries rock particles as sediment; and the lithosphere absorbs both air and water which react with the rocks and change them.

Atmosphere.—The atmosphere is a mixture of gases, chiefly nitrogen, oxygen, argon, carbon dioxide, and water vapor. In addition there are small amounts of helium, ozone, ammonia, sulphurous gases, etc. These substances vary somewhat in amounts, and water vapor varies considerably with the conditions of the atmosphere.

The mass of the atmosphere is about 1/1,200,000 of the mass of the earth. Its pressure is about 14.7 pounds per square inch at the earth's surface. If an open pipe one square inch in cross section could be extended vertically from sea level to a point as high as the atmosphere extends, the weight of the air inside the pipe would be about 14.7 pounds. The density of the air decreases upward, but the air is known to extend about 200 miles above the earth, and probably it extends even farther. The atmosphere is an important geologic agent. It reacts chemically with the rocks and oxidizes them, forming new compounds; it commonly breaks them up into smaller bodies and causes them to disintegrate. This disintegrated rock matter, being more finely divided, is more readily removed by wind and water. Wind—the air in motion—is an important agent of transportation. Moreover, the wind causes the movement of water to form waves and certain ocean currents. Water continually is evaporated into the atmosphere and subsequently is precipitated on the earth as rain. The atmosphere serves also as a thermal blanket which distributes the heat of the sun and tends to prevent its escape from the earth. The dust particles suspended in the atmosphere aid in the diffusion of sunlight. Furthermore, this gaseous envelope which surrounds the earth protects it from the violent bombardment by meteorites. Several million meteorites fall into the earth's atmosphere daily, but most of them are disintegrated by the heat of friction which is generated as they travel through air toward the earth's surface.

Hydrosphere.—The hydrosphere includes the ocean, seas, lakes, rivers, and creeks together with the water that has soaked into the ground and occupies the openings in the lithosphere. The greater part of the water of the hydrosphere is in the ocean. The ocean at places is about 6 miles deep, and its average depth is about 2½ miles. If the water of the ocean were evenly distributed over the entire earth, its depth would be about 2 miles. The surface of the ocean is about 143,000,000 square miles.

Water is an important geologic agent. The rain falls upon the land and gathers into rills which unite to form brooks and creeks. These join to form rivers which, in general, flow to the ocean. The water carries with it rock particles, and little by little the land is worn away. The process of degradation goes on slowly but continuously. Ultimately the continents are reduced, and their material is carried to the sea and deposited so that great beds of sediments are formed from the degraded land. The source of the energy of running water is the sun, which evaporates the water of the sea, forming clouds, from which the water is precipitated as rain upon the earth. Thus the process of construction of new layers of rock goes on along with the destruction of the land. Air and water and vegetation working together break down the rocks and render them incoherent and therefore more readily removed by running water. Where the loose material has not been removed, it constitutes the soil and mantle rock.

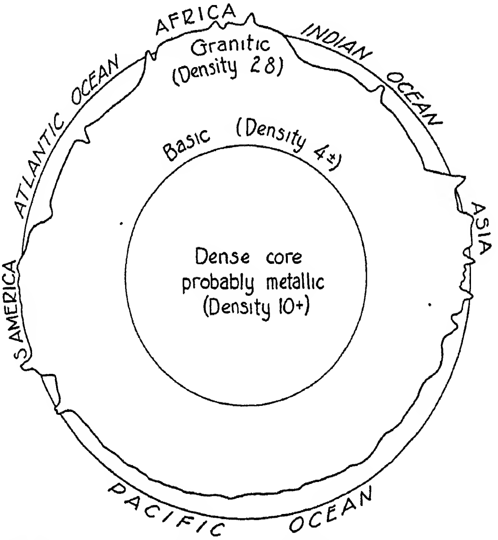

FIG. 4.—Section through the earth at equator showing relief and nature of interior. Relief much exaggerated.

Lithosphere.—The lithosphere is an oblate spheroid flattened at the poles, owing to its rotation. The length of its polar diameter is 7,899.7 miles, and that of its equatorial diameter is 7,926.5 miles, or 26.8 longer than its polar diameter. The equatorial circumference of the earth is 24,902 miles, and the area of its surface is about 197,000,000 square miles. About 143,000,000 square miles lies below sea level, and 54,000,000 square miles is land. Bordering the continents there are areas aggregating about 10,000,000 square miles which are relatively shallow seas (Fig. 4), and which are called continental shelves. The top of Mount Everest of the Himalaya Mountains, north of India, the highest mountain, is about 28,871 feet above the sea; and the greatest known depth of the ocean—35,433 feet—is in the Pacific, off the coast of the Philippine Islands. Thus the relief of the lithosphere is about 12.2 miles. This relief is slight, however, compared to the size of the earth (Fig. 5). It would amount to about 1 foot on a globe with a diameter of 658 feet. A section of the earth's crust true to scale is illustrated by Fig. 5. The section is 1,500 miles long, and the heavy line is wide enough to include all irregularities of the earth's surface from the highest mountain peak to the deepest ocean basin.

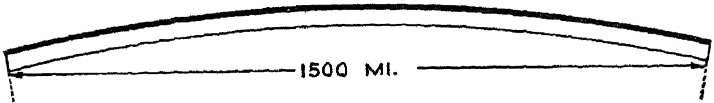

FIG. 5.—Diagram representing a section of the crust of the earth showing the degree of curvature of its surface in a distance of 1,500 miles. The heavy line at the surface is wide enough to embrace all irregularities on the surface of the earth, from the highest peak in the Himalayas to the lowest deep in the ocean basins. The distance between the two curved lines represents a zone about 50 miles thick. (*After J. S. DeLury.*)

In moist climates the lithosphere nearly everywhere is covered with vegetation beneath which are found incoherent materials consisting of sand, clay, and other rock particles mixed with organic matter which has resulted from the decay of vegetation. This thin veneer of the earth's surface is the soil. It passes downward into the subsoil, which is made up of rock fragments. The subsoil contains only a little organic matter. The soil and subsoil constitute the mantle rock. Along streams and in mines and wells it is seen that the mantle rock is underlain by solid rock which is called bedrock. The *crust* of the earth is that part of the solid earth that comes under the observation of man—at the surface, in mines and wells and particularly at places where rocks, once deeply buried after faulting or folding (Figs. 6–10) have been exposed by erosion. This part of the lithosphere is between 5 and 10 miles thick.

Processes Acting on the Lithosphere.—The processes which are now in operation on the surface of the earth include:

1. Gradation:
   a. Degradation, or the wearing down of rocks by water, air, and ice.
   b. Aggradation, or the building up of rock formations by deposition of the degraded material.
2. Diastrophism, or the movements of parts of the solid earth with respect to each other.
3. Vulcanism, or the movements of molten rocks or lavas and the formation of their products.

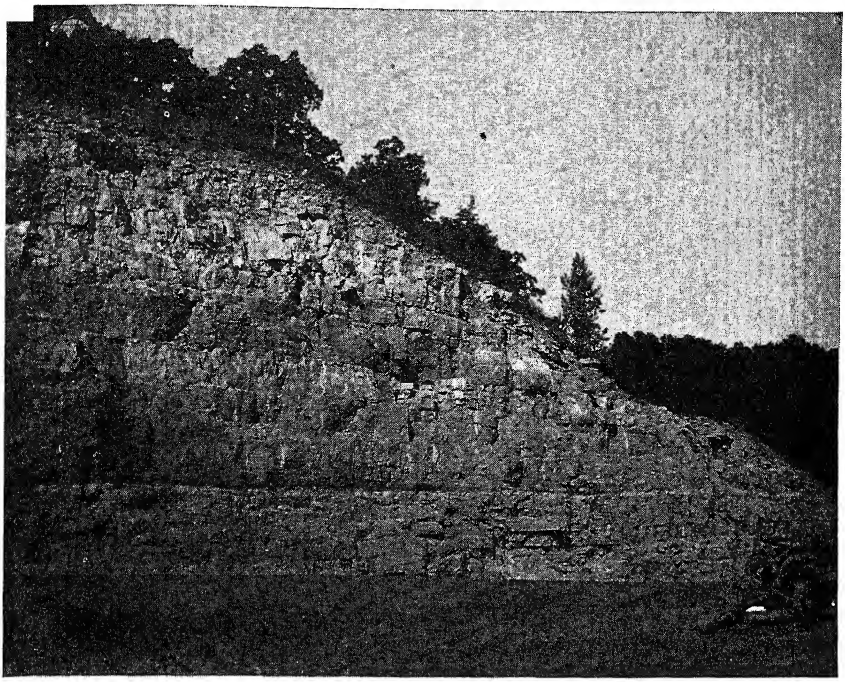

FIG. 6.—Horizontal bedding in sedimentary rocks (Oneota dolomite) along Stockton Hill in the Gilmore Valley at Winona, Minnesota.

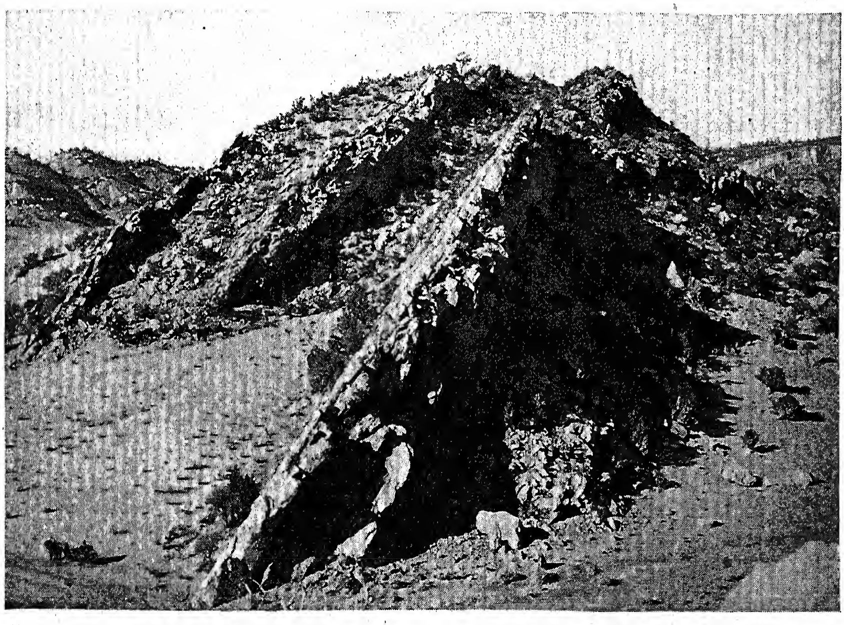

FIG. 7.—Steeply tilted sedimentary rocks near Gallup, New Mexico. (*Photograph by N. H. Darton, U. S. Geol. Survey.*)

Gradation.—Rocks exposed at the surface of the earth are continually subject to gradation. The waves beat upon the coasts and wear away the land. The rain falls upon the land, and the water flows to the creeks and rivers and thence to lakes and seas. The water carries with it particles of rock, and these finally are dropped in lakes, along rivers, and in the ocean, where they form beds of sand and mud. The wind carries rock particles and rolls them along the surface of the earth, depositing them as sand dunes. Ice in motion carries rock fragments; and when the ice melts, these are deposited. The process of gradation is generally slow, but it is continuous, and in long periods of time its results are very great.

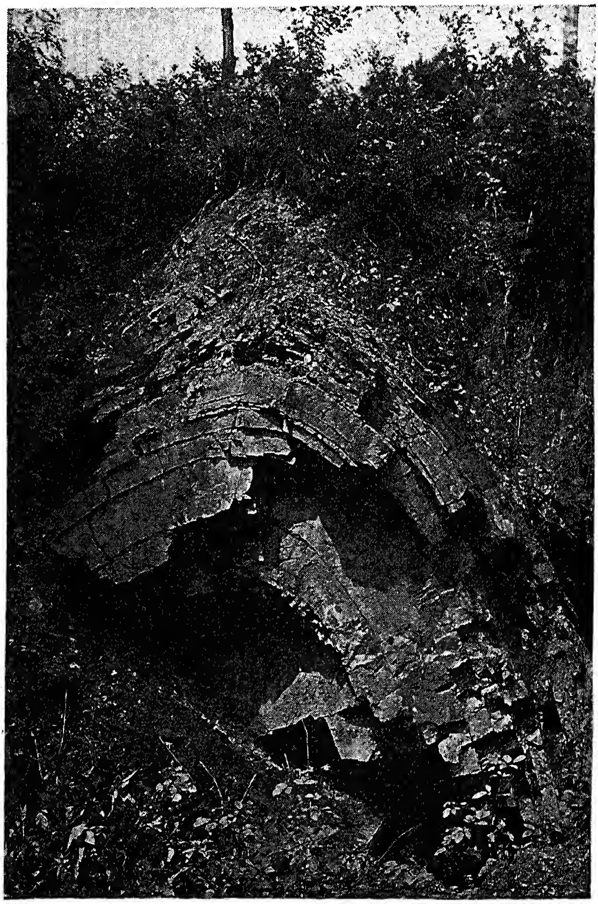

FIG. 8.—Sedimentary beds arched into a fold (anticline); near Lead, Black Hills, South Dakota. (*Photograph by A. J. Tieje.*)

Diastrophism.—Diastrophism includes all movements of solid parts of the earth with respect to other parts. At many places rocks that contain the remains of sea animals are found high above the sea. Such rocks have been raised by diastrophism. A study of the rocks and their relations to each other shows that large areas of the earth's surface have been submerged below sea level and elevated above sea level many times. Diastrophism commonly determines the nature of gradation. Above sea level the dominant process is degradation; below sea level the dominant process is aggradation. When sediments are laid down in the sea, they are nearly flat lying (Fig. 6); if they are found to be folded or on edge (Fig. 7), the inference is warranted that they have been disturbed by diastrophic movements. Degradation tears down the rocks and removes their particles to the sea where new beds are formed from them. By diastrophism these beds are raised above the sea and exposed again to degradation. By diastrophism beds are folded (Fig. 8), and fractured. Where fractures cut the beds and the block on one side is moved with respect to the block on the other side of the fracture, the fissure becomes a fault (Fig. 9).

Fig. 9.—Sedimentary beds displaced along a fracture producing a fault.

Vulcanism.—Vulcanism includes all of the phenomena that are connected with molten rock matter and its movements. Large parts of the earth's surface are made up of rocks that have solidified from the molten state. Such rocks are igneous rocks. Some of them are extrusive rocks that have been ejected from the throats of volcanoes; others, the intrusive rocks, solidified at depths. The deep-seated intrusive igneous rocks were

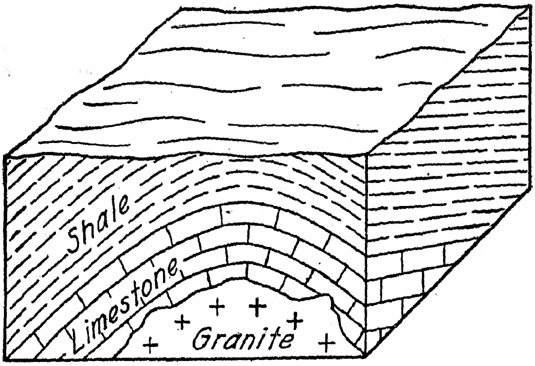

FIG. 10A.—Diagram showing sedimentary rocks intruded by granite. The granite was formed at great depth and is not exposed at the surface. The rocks above the granite are folded and arched up.

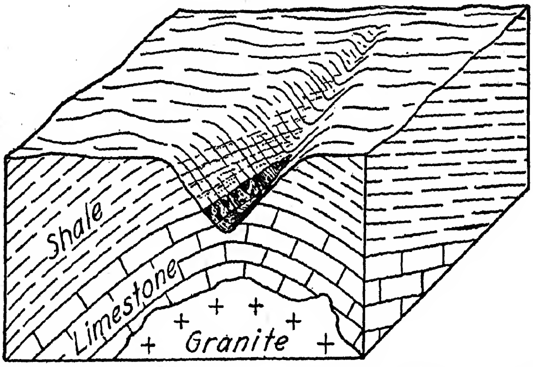

FIG. 10B.—Diagram showing sedimentary rocks intruded by granite. A stream has cut into the limestone, but the granite is not exposed.

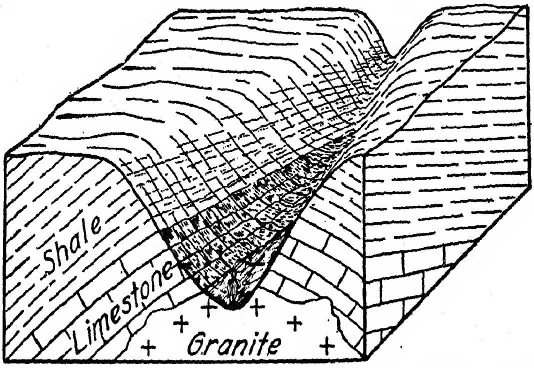

FIG. 10C.—Diagram showing sedimentary rocks intruded by granite at great depths. A stream has cut a valley through the sedimentary rocks, and the granite is now exposed to the view of man.

not exposed at the surface of the earth until erosion had removed the material that covered them (Fig. 10 *A*,*B*,*C*). When igneous rocks are exposed above the sea at the surface of the earth, they are attacked by air and water and are broken down. Their fragments, together with other material, are carried away by streams and are deposited to form sands and muds.

Methods of Study.—The geologic history of the earth should be interpreted in the light of the processes that are observed to be operating on the earth today. These processes—gradation, diastrophism, and vulcanism—are studied, and their results are noted as a means of interpreting the geologic past. This principle was developed chiefly by the British geologist Lyell.

The methods of geology are very simple if the steps are taken singly. If two sheets of paper are dropped one at a time, it is evident that the first one dropped will lie below the second. If these sheets are not disturbed and if they are found many years later, it is obvious that the one at the bottom was laid down before the one above it. Similarly the relative ages of a group of bedded rocks may be determined. In the rocks are found the remains of plants and animals that were living while the rocks were being formed. The remains in one group of beds are different from those of other beds of the same group of rocks, but they may resemble the remains of organisms that are found in beds located far away. Thus a bed carries a "label" that states its age. By studying hundreds of sections of rocks and mapping them, the extent of the area over which a bed was deposited may be learned, and thus it is possible to chart the ancient sea in which a bed was deposited many ages ago.

Where sediments are laid down in water, they are nearly flat (Fig. 6); where beds formed from sediments have been disturbed by diastrophism, they are found to be tilted and folded (Fig. 8). Where an igneous mass intrudes a group of beds, it is known to be later than the beds intruded. Thus by studying large areas—the beds, their attitudes, and their relations to each other and to igneous bodies—it is possible to ascertain the times at which the major events have taken place in the area containing the rocks and to interpret the geological history of the area containing them.

Most geological processes are very slow, and geologic time is exceedingly long, especially when measured by the scale of human history. With an increase in our knowledge of earth's history there comes an extension of our conception of geologic time. Processes and forces which seem to have but slight effect when observed from day to day are capable of producing tremendous results when continued over long periods of geologic time. The geologist looks upon the duration of geologic time as comparable to the vast expanses of space recognized by the astronomer.

Main Divisions of Geologic Time
(In part from Schuchert and Mather)

Eras	Periods	Outstanding physical events	Life development
Cenozoic	(Quaternary)		
	Recent	Postglacial changes	Age of mental dominance
	Pleistocene	The Great Ice Age	Advent of man
	(Tertiary)	Formation of Coast Range	
	Pliocene	Formation of the Alps and many other mountain chains	Primitive horse and other ungulates
	Miocene		
	Oligocene	Extensive volcanic activity in western United States	
	Eocene		First placental mammals
Mesozoic		LARAMIDE REVOLUTION	
	Cretaceous	Early folding to form Rocky Mountains	
			Extinction of dinosaurs
			Climax of reptiles on land, air, and sea
	Comanchean	Central cordilleran disturbance	First flowering plants
	Jurassic	Beginning of Sierra Nevada uplift	First birds
	Triassic	Extensive volcanic activity in New England, Pennsylvania, and New Jersey	First dinosaurs
Paleozoic		APPALACHIAN REVOLUTION	
	Permian	Folding to form Appalachian Mountains, glaciation	Rise of reptiles
	Pennsylvanian	Extensive coal-forming swamps	Large non-flowering plants
	Mississippian	Paleozoic Alps	First land vertebrates
	Devonian	Acadian Mountains	Age of fishes
	Silurian	Caledonian uplift	Rise of land plants
	Ordovician	Taconic Mountains	First known fishes
	Cambrian	Green Mountains disturbance	Age of invertebrate dominance
Proterozoic		PENEPLANATION, GLACIATION	
		LIPALIAN INTERVAL	
		KILLARNEY REVOLUTION	
	Keweenawan	Extensive lava flows	Scanty record of primitive plants and animals
	Huronian	Ocoee Mountains (Schuchert)	
	(Animikian)	Oldest evidence of glaciation	
	Algoman	Folding connected with igneous intrusions	
	Timiskamian		
Archeozoic		LAURENTIAN REVOLUTION	
	Laurentian		
	Keewatin	Rocks much altered and history obscured	Primitive life

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CHAPTER II

NATURE OF MATERIALS THAT CONSTITUTE THE EARTH

All matter is made of atoms. Different kinds of atoms constitute the chemical elements, and all known substances are nothing more than combinations of these elements. Only 93 different elements are known, and of these a considerable number are found only rarely in the materials of the earth. In fact eight different kinds make up about 98 per cent by weight of the earth's crust (Fig. 11). Examples of different kinds of elements are the metals iron, copper, silver, etc., and well-known gases such as oxygen, hydrogen, and chlorine, together with less widely known substances like phosphorus, strontium, and cobalt. Atoms are exceedingly small, the average diameter being about one hundred-millionth of an inch. These minute particles are believed to consist of an electropositive nucleus,[1] attended by one or more electronegative units known as electrons. Thus an atom of hydrogen is composed of one proton and one electron, whereas in an atom of carbon the nucleus is capable of gathering six electrons about it. Atoms show also marked differences in weight.

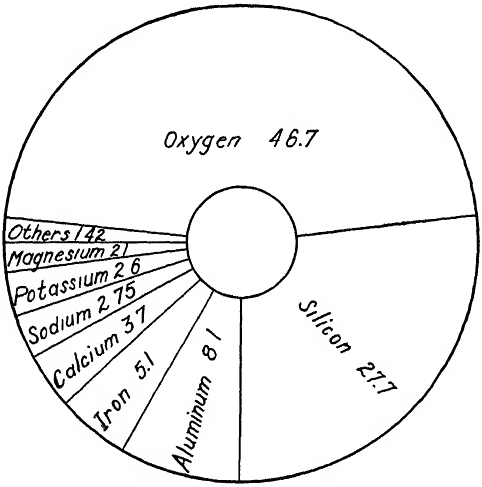

FIG. 11.—Diagram showing the relative abundance of the most common elements of the earth's crust. All figures indicate percentages. (*After Clark and Washington.*)

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[1] The nucleus of hydrogen consists of one proton. The nuclei of other atoms contain both protons and electrons, the protons being in excess.

The heaviest atom is that of uranium, which is about two hundred thirty times as heavy as one atom of hydrogen, which is the lightest element.

Some kinds of atoms have a tendency to stick together in pairs or in groups of three or more, either with atoms of their own type or with other kinds of atoms. The products of such unions are molecules. A molecule may be defined as the smallest particle of a certain kind of substance which under ordinary conditions may exist as a unit and still retain the properties of the original material. Thus there are molecules of salt, molecules of sugar, and molecules which at ordinary temperatures are gases, such as molecules of oxygen and nitrogen. An oxygen molecule differs from one of salt in that it is composed of a union of two like

ELEMENTS FOUND IN THE EARTH'S CRUST AND PERCENTAGE OF CRUST FORMED BY EACH ELEMENT

(After Clark and Washington)

Element	Average of 10 miles of crust[1]	Average of igneous rocks, 10-mile crust
Oxygen	46.71	46.59
Silicon	27.69	27.72
Aluminum	8.07	8.13
Iron	5.05	5.01
Calcium	3.65	3.63
Sodium	2.75	2.85
Potassium	2.58	2.60
Magnesium	2.08	2.09
Hydrogen	0.14	0.13
Titanium	0.62	0.63
Chlorine	0.045	0.048
Phosphorus	0.13	0.13
Carbon	0.094	0.032
Manganese	0.09	0.10
Sulphur	0.052	0.052
Barium	0.050	0.050
Chromium	0.035	0.037
Nitrogen		
Fluorine	0.029	0.03
Zirconium	0.025	0.026
Nickel	0.019	0.020
Strontium	0.018	0.019
Vanadium	0.016	0.017
Cerium, yttrium	0.014	0.015
Copper	0.010	0.010
Others	0.033	0.034
	100.000	100.000

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[1] Includes both igneous and sedimentary rocks.

oxygen atoms, whereas salt is composed of one sodium atom united with one atom of chlorine. Most molecules are very small. The number in a cubic inch of air at ordinary temperature and pressure is represented by the number 6 followed by 19 ciphers (Bazzoni)—a number so immense that it is beyond ordinary comprehension. Even though the number of molecules is enormous, they may not be closely spaced. In all gases they move rapidly to and fro in an independent existence. In liquids, however, the molecules are more closely spaced and are always in contact with one another even though their motion causes them to shift, turn, vibrate, and change partners continually. The molecules of water are always in motion, but their movement is not sufficient to cause them to break away entirely from one another. That is why at ordinary temperatures water is a liquid rather than a gas. In still other substances the

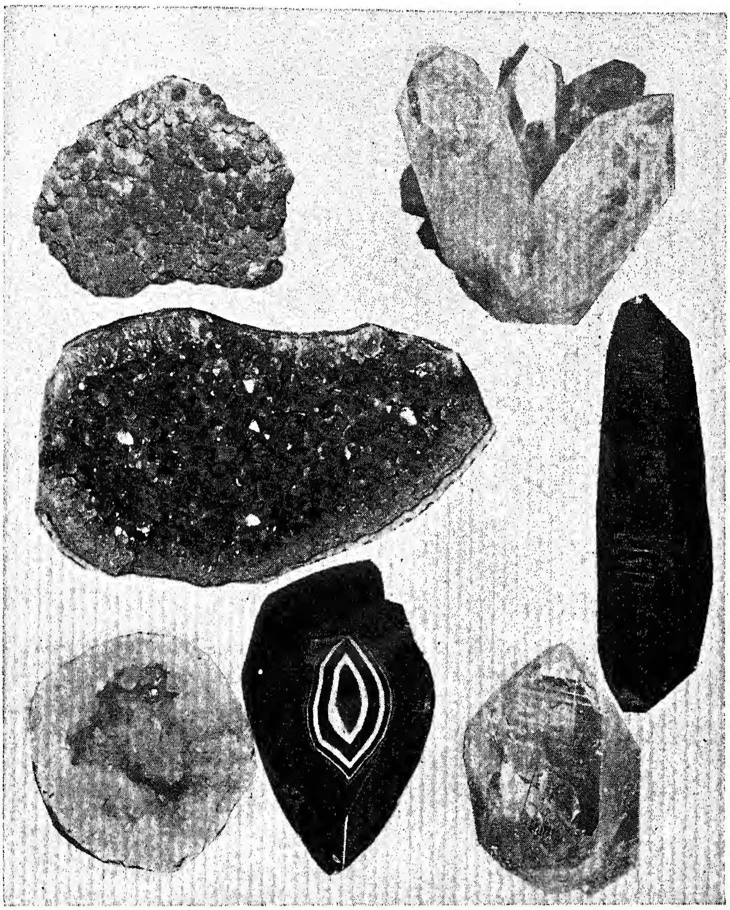

Fig. 12.—The mineral quartz, showing various physical forms. (*Courtesy Ward's Natural Science Establishment.*)

mutual attractions of molecules or atoms counterbalance the effects of unorganized motion, and each molecule or atom is tied to its neighbors and a solid is formed. These three phases of matter—gaseous, liquid, and solid—are exemplified on earth as the air, the water, and the land respectively. The probable composition of the interior of the earth is illustrated by Fig. 4.

MINERALS

A mineral is a natural compound with a chemical composition and physical properties that are either uniform or variable within definite limits. Minerals are formed by precipitation from various kinds of solutions. Most of them are solids that have a characteristic crystalline structure due to the arrangement of the molecules of which they are composed. These compounds represent combinations of the chemical elements that have been identified in the earth. Eight elements, as stated, make up about 98 per cent of the earth's crust (Fig. 11). Oxygen is the most abundant, and the seven other elements unite with oxygen to make up many of the common minerals. The most fundamental combination of these elements is their union with oxygen to form oxides. When silicon unites with oxygen, silicon dioxide is formed, which unites with water and forms acids. The six other elements unite with oxygen and

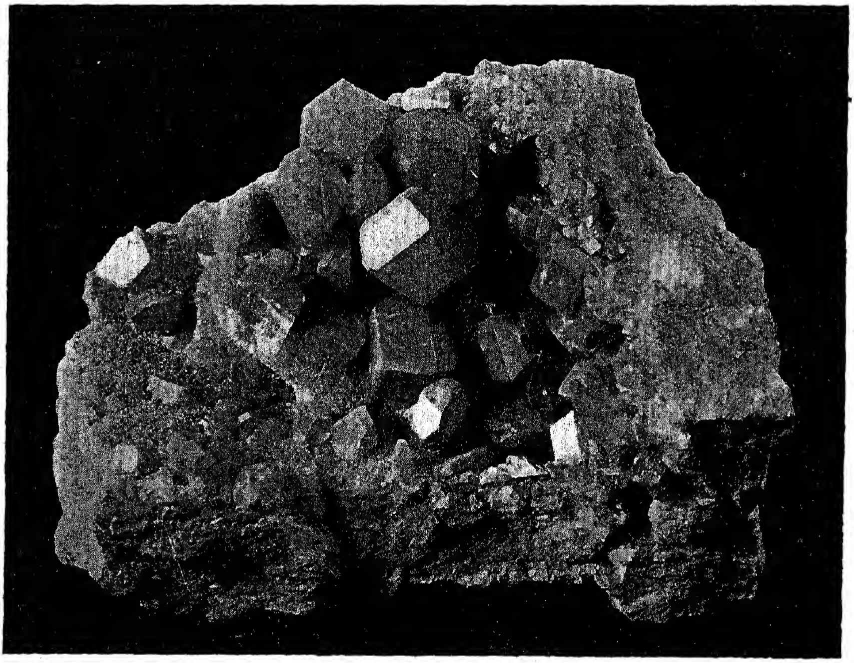

FIG. 13.—Crystals of sulphur. (*Courtesy American Museum of Natural History.*)

water to form bases. The acids and bases combine to form silicates, which are the most abundant compounds in the earth's crust. These silicates, together with free silica and free iron oxide (magnetite), form the minerals which are the chief constituents of igneous rocks. If silica is abundant in a rock, the rock is said to be silicic or acidic. If the bases are abundant, the rock is said to be basic.

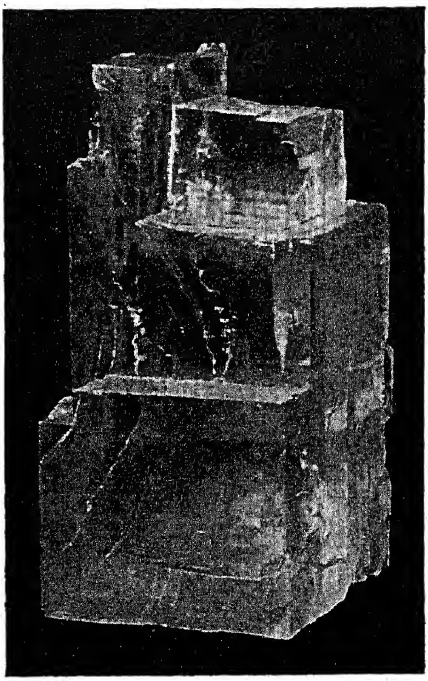

FIG. 14.—A crystal of halite showing cubic cleavage. (*Courtesy Ward's Natural Science Establishment.*)

Nearly all of the minerals are crystalline (Figs. 12, 13, 14), and therefore the earth's crust is essentially crystalline. Crystals are organized in such a way that their outlines commonly have definite symmetrical forms such as the forms frequently seen in snowflakes. Their substance also has a definite geometrical structure due to the organization of the material of which they are made. Crystalline materials have nearly constant properties, and it is possible to recognize and identify most of them by their forms, appearance, hardness, and other characteristics. Although crystals are commonly identified by form, the effect of crystallization extends beyond the surface to all parts of the substance. A minute fragment of a certain crystal shows a definite characteristic cleavage. It affects light in a certain way, and hence nearly any fragment may be identified under the petrographic microscope by observing its effect on light.

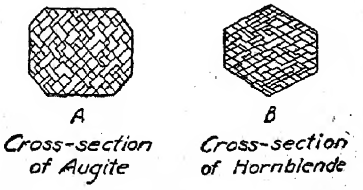

FIG. 15.—*A*, cross section of a crystal of augite showing characteristic cleavage; *B*, cross section of a crystal of hornblende (amphibole) showing characteristic cleavage. (*Redrawn from Miers.*)

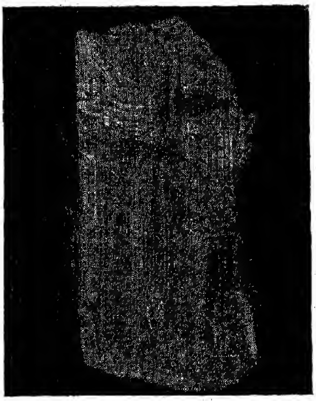

FIG. 16.—Asbestos, a fibrous mineral that breaks in minute threads.

Hardness.—The hardness of a mineral, or its resistance to abrasion, is a fairly constant quality. The hardness generally is designated after comparing it with the Mohs' scale, in which 10 minerals are arranged in increasing hardness as follows: (1) talc, (2) gypsum, (3) calcite, (4) fluorite, (5) apatite, (6) orthoclase feldspar, (7) quartz, (8) topaz, (9) sapphire (corundum), (10) diamond.

The minerals in the part of the scale below 6 can be scratched with a knife. In the field a small-bladed penknife kept sharp at the point is found very useful, since with a little practice one may estimate closely the hardness and at the same time observe the streak. A few minerals possess magnetism, and their particles are easily picked up by a small magnet. If the knife blade is magnetized, it is more useful.

Weight.—The specific gravity or density of a mineral is its weight compared with a volume of water equal to that of the mineral. The specific gravity of a small piece is easily determined by using a spring balance, first weighing it in air and then in a small pan suspended in a beaker of water. Pure specimens of the same mineral generally have approximately the same weight or specific gravity, and so the determination of its specific gravity is a useful aid in the identification of a mineral.

Tenacity.—Certain minerals which powder easily are brittle. Others, like gold, are malleable and can be hammered into thin sheets. Still others, like horn silver, are sectile; they cut like cheese. A mineral that bends yet does not resume its original shape when pressure is released is said to be flexible. Chlorite is an example. An elastic mineral (mica) after being bent will resume its original shape.

Other Properties.—In the preceding statements emphasis is laid upon those properties of minerals that can be used to distinguish them outside the laboratory. With a blowpipe and the use of a few simple reagents, chemical tests may be made, and mineral powders or fine-grained mixtures may be investigated. The minerals differ in fusibility and in their behavior with reagents. The silicates often may be identified in the field, but exact determinations are most easily made with the microscope. The minerals or rocks are ground into thin slices, and light is passed through them. Very exact determinations can be made, because the effect on light of every transparent mineral differs from that of every other one. Opaque minerals, like the sulphides of the metals, are easily studied under the microscope by reflected light.

Isomorphous Mixtures.—The mineral sphalerite, zinc sulphide (ZnS), commonly contains some iron (Fe). The light-yellow species is practically free from iron. If a little iron is present, its color is brown; if more iron is present, it is black. Its formula

MINERAL COMPOSITION OF AVERAGE IGNEOUS AND SEDIMENTARY ROCKS
(*Clark and Washington*)

Igneous rocks		Sedimentary rocks		
Minerals	Per cent	Minerals	Shale, per cent	Sandstone, per cent
Quartz	12.	Quartz	22.3	66.8
Feldspars	59.5	Feldspars	30 0	11.5
Pyroxenes and amphiboles	16.8	"Clay"	25.0	6.6
Micas	3.8	Limonite	5.6	1.8
Other minerals	7.9	Carbonates	5.7	11.1
		Other minerals	11.4	2 2
Total	100.0	Total	100.0	100.0

would be written (Zn, Fe)S to show that some of its molecules are ZnS and others, FeS. The Fe is there in place of Zn, yet the crystals of sphalerite may have the same form and essentially the same structure, and any particle of the dark sphalerite would show, on analysis, the same composition as any other part of it. This mixture, in which one molecule takes the place of another without changing the crystal form, is known as an isomorphous mixture. An isomorphous group of compounds is one in which a series of compounds have closely related chemical composition and nearly similar crystals. Thus calcite (calcium carbonate), magnesite (magnesium carbonate), and siderite (iron carbonate) all crystallize in the same crystal system. They form isomorphous mixtures in nearly all proportions. Many calcites contain some magnesium carbonate; others contain iron carbonate; and still others contain both. The minerals that are isomorphous mixtures constitute a large part of the earth.

COMMON MINERALS

Although many hundreds of minerals have been identified, only a few of them are of common occurrence. The earth's crust is made up chiefly of seven. The estimates stated in the table on page 21 show approximately the minerals which compose the average igneous rocks, the shales, and the sandstones. The limestones are essentially calcium and magnesium carbonate rocks with some quartz, clay, and iron minerals, and the metamorphic rocks are chemically similar to igneous and sedimentary rocks.

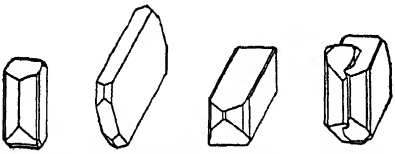

FIG. 17.—Drawings of crystals of orthoclase feldspar. (*Niggli.*)

Feldspars.—The feldspars (Fig. 17) make up about half the rocks of the earth's crust. They are generally light colored and are characterized by two good cleavages. In orthoclase these make angles of 90 degrees; hence the name.[1] Albite and anorthite have cleavages that make angles of about 86 degrees; hence they are termed plagioclases.[2]

Orthoclase feldspar, potassium aluminum silicate,[3] is commonly pink or flesh-colored. Albite, sodium aluminum silicate, is usually white; and anorthite, calcium aluminum silicate, is commonly gray-green.

Quartz.—Quartz,[4] silicon dioxide, next to feldspar the most abundant material of the earth's crust, forms six-sided crystals (Fig. 12) and is

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[1] Orthoclase, Greek, *orthos*, straight; *klao*, I break.

[2] Plagioclase, Greek, *plagios*, oblique; *klao*, I break.

[3] The formulae of the feldspars are: orthoclase, $K_2O.Al_2O_3.6SiO_2$, hardness 6; albite, $Na_2O.Al_2O_3.6SiO_2$, hardness 6.5; anorthite, $CaO.Al_2O_3.2SiO_2$, hardness 6.5. Albite and anorthite molecules form a series of isomorphous mixtures or compounds: albite, oligoclase, labradorite, bytownite, anorthite. Starting with albite the sodium decreases, and calcium increases to anorthite. The physical properties of the isomorphous crystals vary with the composition, particularly the effect on light, as the light passes through the crystals under the microscope. The albite-anorthite series are generally striated, and with a small lens the striations often may be observed as very fine parallel lines closely set like ruled lines on a sheet of paper.

[4] The formula of quartz is SiO_2. Quartz is harder than feldspar and cannot be scratched with a knife. Its crystals scratch glass. Its hardness is 7.

colorless or white when pure, although it is commonly tinted. It has no cleavage, and hence it rarely presents flat surfaces when broken. In a granite in which the light-colored minerals are quartz and feldspar, the quartz may be distinguished from the feldspar by turning the rock so as to get a reflection from the cleavage planes of the feldspar. Because it has no good cleavage the quartz breaks like glass.

Pyroxenes.—The pyroxenes (Fig. 18) constitute an important group of minerals that are generally recognized by their stout crystals and their two good cleavages (Fig. 15*A*) almost at right angles to each other (87 and 93 degrees). Most pyroxenes are green or dark colored, particularly those that contain much iron. Pyroxenes are calcium, magnesium, and iron silicates, and some have very complicated formulae.[1]

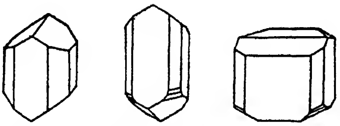

FIG. 18.—Drawings of crystals of pyroxene (augite).

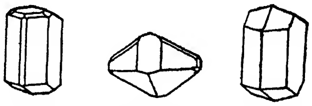

FIG. 19.—Drawings of crystals of amphiboles.

Hornblendes.—The hornblendes, or amphiboles (Fig. 19), constitute an important group of minerals which may occur as stout crystals, but more generally they are long-bladed or fibrous ones, green to black in color. The hornblendes are calcium, magnesium, iron, and aluminum silicates and generally have complicated formulae.[2] The cleavage angles are 125 and 55 degrees (Fig. 15*B*).

Micas.—The micas are distinguished from other minerals by their perfect cleavage, which makes it possible to separate them into extremely thin sheets. Muscovite, white mica (hydrous potassium aluminum silicate), is the variety used for stove windows. Biotite, black mica, which resembles muscovite except in color, has nearly the same chemical composition but contains some iron or magnesium or both.[3]

Olivine.—Olivine,[4] magnesium silicate, is found in many basic rocks and usually occurs in stout crystals. It is glassy like quartz but is generally olive green or yellow. Its grain in a rock often resembles that of granulated sugar. Clear varieties are used for gems.

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[1] Diopside, $CaMg(SiO_3)_2$; hypersthene, $(Mg,Fe)SiO_3$; augite, like diopside but with aluminum and iron. As a rule, pyroxenes have a dull luster, and this is an aid in distinguishing them from amphiboles (hornblende), which commonly have a silky sheen.

[2] Tremolite, $Ca_2Mg_5Si_8O_{22}(OH)_2$; actinolite, $Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2$; common hornblende contains also alumina and soda. They commonly have a glittering, silky sheen, whereas pyroxenes have a duller luster.

[3] Muscovite, $KAl_2(AlSi_3O_{10})(OH)_2$; biotite, $K(Mg,Fe)_3(AlSi_3O_{10})(OH)_2$.

[4] Olivine, $(Mg, Fe)_2SiO_4$. In many olivines the iron takes the place of part of the magnesium.

Garnet.—The garnets[1] are iron, calcium, or magnesium silicates, usually red or brown with vitreous luster. They have no good cleavage and break like quartz or like glass.

Chlorite.—Chlorites are silicates of aluminum containing magnesium, iron, and hydrogen. They are green to dark green, and the crystals resemble mica because they have one excellent cleavage. Unlike mica the cleavage plates are not elastic, and when bent they do not resume their original shape on release. Chemically chlorite is much like biotite but with no potassium, less silica, and more water.

Kaolinite.—Kaolinite,[2] hydrous aluminum silicate, is a soft and usually light-colored mineral that occurs in minute particles. It is a constituent of many clays and shales. It feels greasy between the fingers.

Calcite, Dolomite, and Siderite.—Calcite, calcium carbonate; dolomite, calcium magnesium carbonate; and siderite, iron carbonate,[3] are all soft minerals that are characterized by good cleavage. Calcite is the chief constituent of limestone, and dolomite is present in dolomitic limestone. Siderite is found in sedimentary rocks and is a common ore of iron.

Magnetite.—Magnetite,[4] black iron oxide, is a dark, heavy magnetic mineral that is present in small amounts in most igneous rocks. It is brittle, has no good cleavage, and is too hard to be scratched with a knife; streak black; an ore of iron.

Hematite.—Hematite,[5] red iron oxide, is the chief ore of iron. It has a red streak, like rouge or red paint.

ROCKS

Minerals are combined to form rocks. It has been shown that the elements combine in various proportions to form minerals or to form molecules that unite to form the isomorphous mixtures. Rocks, on the other hand, are mixtures of various minerals in almost all proportions. Some mixtures, however, are much more common than others.

The principal rocks are igneous, sedimentary, and metamorphic. Igneous rocks are formed by the consolidation of molten matter. Figure 20 shows a granite, an igneous rock formed at great depths. Sedimentary rocks are formed from fragments of igneous rocks, from metamorphic

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[1] Almandite, common red garnet, $Fe_3Al_2(SiO_4)_3$; pyrope garnet, $Mg_3Al_2(SiO_4)_3$; grossularite garnet, $Ca_3Al_2(SiO_4)_3$; andradite garnet, $Ca_3Fe_2(SiO_4)_3$. Garnets are about as hard as quartz.

[2] Kaolinite, $Al_2Si_2O_5(OH)_4$.

[3] Calcite, $CaCO_3$, hardness 3; dolomite, $CaMg(CO_3)_2$, hardness 3.5 to 4; siderite, $FeCO_3$, hardness 3.5 to 4.

[4] Magnetite, Fe_3O_4 or $FeO.Fe_2O_3$.

[5] Hematite, Fe_2O_3.

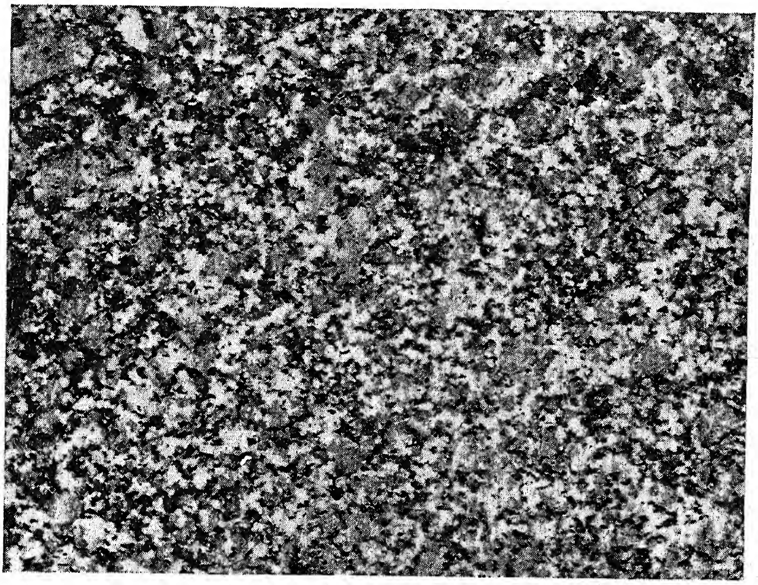

FIG. 20.—Granite, an igneous rock.

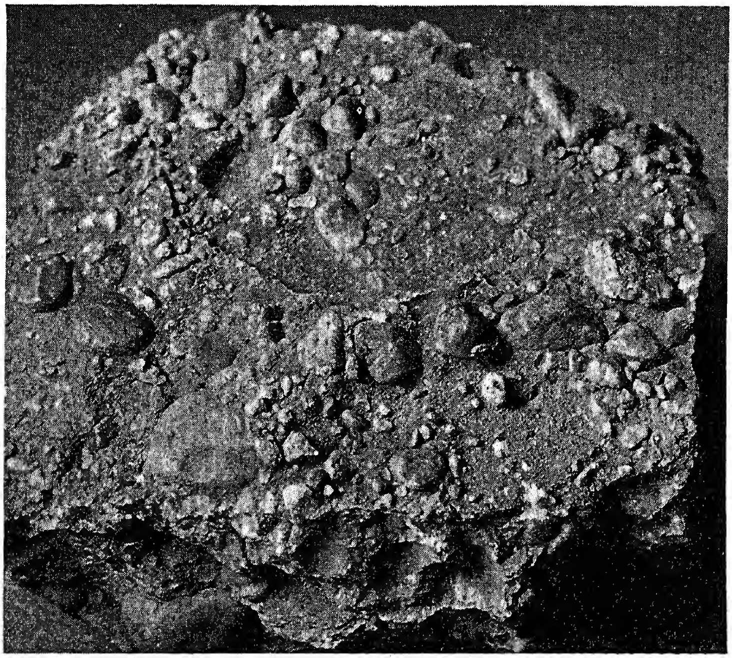

FIG. 21.—Conglomerate, a sedimentary rock.

rocks, or from older sedimentary rocks. Figure 21 shows a conglomerate which is formed from fragments of older rocks. Metamorphic rocks are the changed products of igneous or sedimentary rocks or other metamorphic rocks. Figure 22 shows a gneiss which is a metamorphic rock formed by pressure from granite.

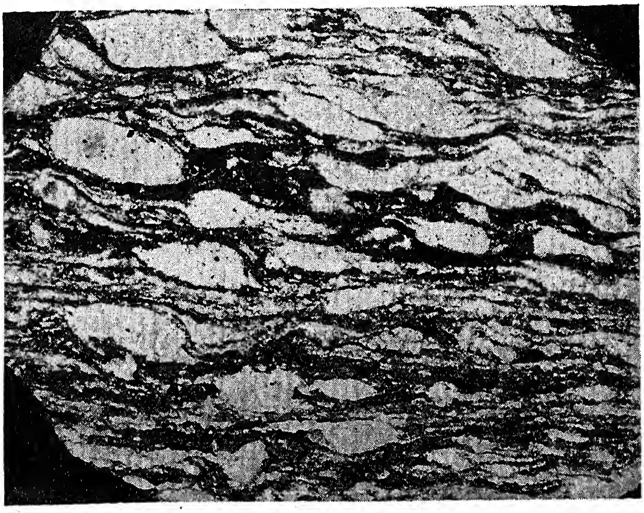

FIG. 22.—Gneiss, a metamorphic rock. A granite is metamorphosed by pressure to form a gneiss. The large white crystals are feldspars; the bands between them are quartz, biotite (black mica), and feldspar.

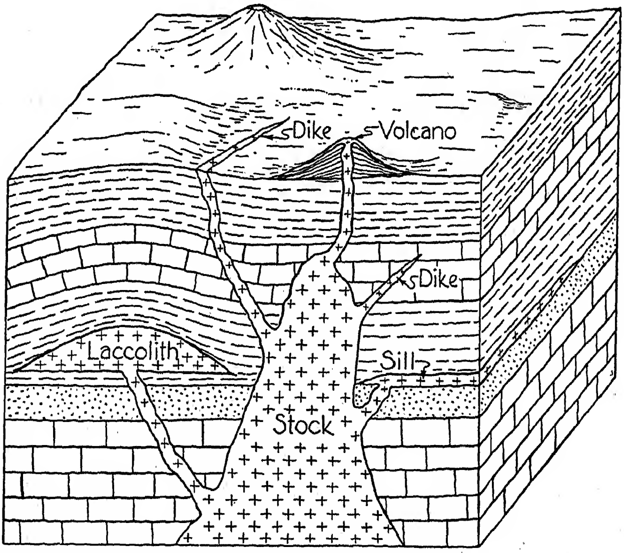

FIG. 23.—Diagram showing structural relations of igneous masses.

Igneous Rocks.—The molten matter which solidifies to form the igneous rocks is termed the magma. The character of the igneous rock depends not only upon the chemical composition of the magma that formed it but also upon the conditions that prevailed when the magma cooled. If it is thrown out upon the surface, it is a lava, or extruded rock; if it is thrust into the earth's crust and does not reach the surface, it is an intrusive rock (Fig. 23). There are certain textures shown by igneous rocks which depend largely on the conditions of their formation. Thus the deep-seated rocks are composed of crystals or grains and are said to be granular (Fig. 20). Granular rocks were formed far below the surface, and for that reason they are exposed to the observation of man only where the surface rocks above them have been washed away (Fig. 10). Volcanoes are vents through which rocks issue, either molten rocks or fragments broken up chiefly by the explosions of gases. Magmas contain dissolved gases; when the magma is thrown out, the gas expands, owing to release of pressure, and a bubble or blowhole forms. These holes are called vesicles (Fig. 24).

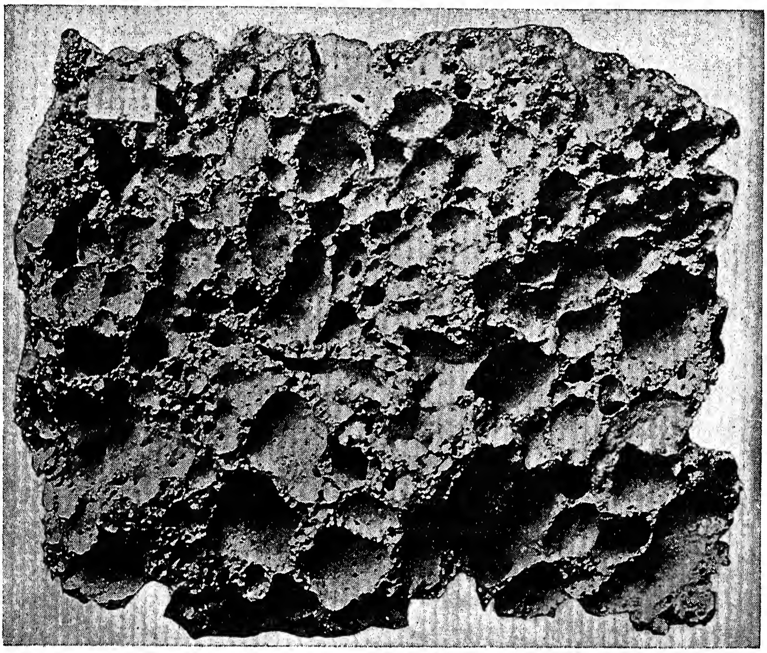

Fig. 24.—Photograph of scoria showing vesicles or cavities formed by expansion of gases in lava during cooling. (*Courtesy Henry Holt & Company.*)

When a magma rising in a crack or fissure hardens, it forms a dike (Fig. 23). A dike is commonly tabular; that is, it is shaped like a tablet, long in two dimensions and short in the third. Magmas intruded between layers of sedimentary rocks form sheets or sills. Many of these are nearly flat, like the sill of a doorway (from which the name is taken), but they may be tilted at high angles. Sills are similar to dikes, but they lie between beds. A thick sill that bows up the rocks above it becomes a laccolith (Fig. 23). A very large, irregular, deep-seated intrusion is called a batholith (Fig. 25). A batholith is long in each of its three dimensions.

Sedimentary Rocks.—Sedimentary rocks (Figs. 6, 7) are derived from the waste products of older rocks. In the presence of air and water, rocks weather or disintegrate. By weathering, some minerals are dissolved and others are released from the rocks and broken up. Rock waste is picked up by rills and carried to the brooks, thence to the creeks, and to rivers. Finer materials are carried in suspension; the coarser are rolled along the stream beds. Much material is dissolved from rocks and carried to the streams in solution. The materials—coarse, fine, and dissolved—generally find their way to the lakes and seas, where they are deposited. Along the shores of the seas they are mixed with materials cut away from the cliffs. The whole is reassorted. The coarser material is deposited near shore as gravel and farther out as sand; still farther mud is deposited; and, farther still, calcareous matter.[1] The belts of sediments are thus rudely parallel to the shores. They are rarely pure, however, for gravels generally contain sand, the sands generally contain mud or clay, the muds contain much fine sand and generally some calcareous matter, and calcareous rocks contain both clay and sand. Because the conditions of their formation are not uniform, muddy sands may alternate with muds. Likewise, muds are deposited with the calcareous rocks. This alternation of material, or layering, is called stratification (Fig. 9) and it is a common feature of essentially all sedimentary rocks. When the sediments are deeply buried below younger beds,

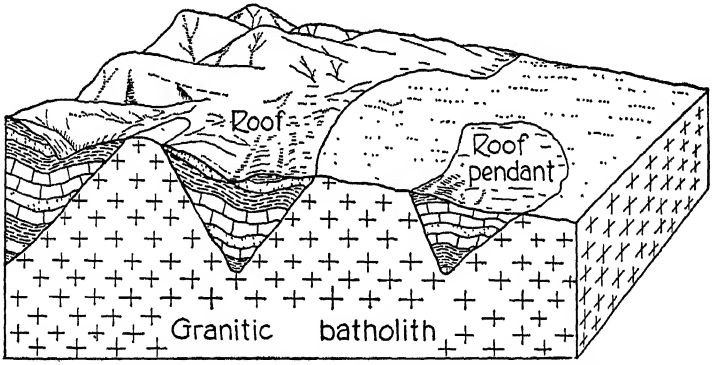

FIG. 25.—Diagram showing a partially exposed batholith.

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[1] This is the general rule. At places sand, mud, and calcareous matter are deposited near shore, particularly in protected areas where the water is not rough. By some the term *marl* is applied to incoherent, non-marine, calcareous matter but not to marine calcareous mud.

they become consolidated by pressure. Some water is squeezed out of them, and by cementation they become coherent; gravels become conglomerates, sands become sandstone, muds become shales, and calcareous oozes become limestones.

The sedimentary rocks commonly contain fossils or the remains of plants and animals that lived while the rocks were being deposited. Some fossils are found only in the beds that were formed at certain times, and such fossils serve as markers that show the relative ages of the beds containing them.

Metamorphic Rocks.—The metamorphic rocks are igneous or sedimentary rocks that have been greatly changed since they were formed. The changes are due to pressure, heat, recrystallization, and cementation.

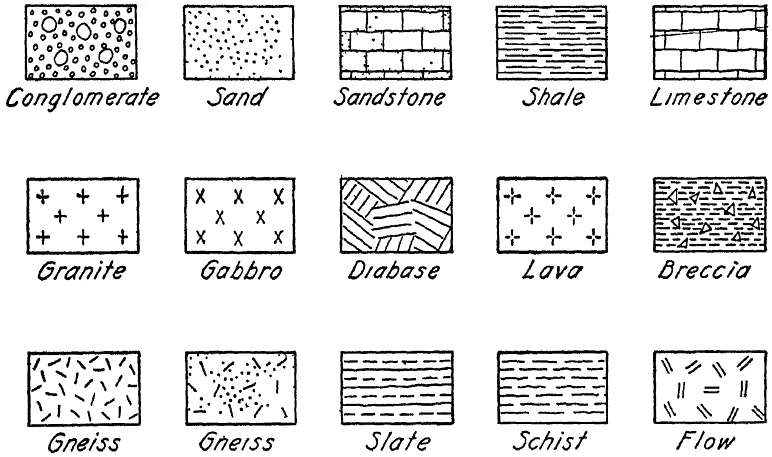

Fig. 26.—Conventional symbols used in geological cross sections. Those shown in upper row and the symbols for slate and schist are very generally used to show rocks listed. The other symbols are used with less uniformity.

Gradually the changes become so great that the original rocks are profoundly altered. Their textures are transformed, and new minerals are developed. Figure 22 shows a granite metamorphosed to form gneiss. In one type of metamorphism recrystallization is the chief process. Thus a calcareous sediment which on slight recrystallization and induration becomes a limestone will by thorough recrystallization become a marble. A marble has the same composition as the limestone, but the calcium carbonate crystals are larger. A sand which on consolidation becomes a sandstone by metamorphism will be thoroughly cemented by the deposition of silica around the sand grains, and thus it becomes quartzite. A mud upon induration becomes a shale; under strong pressure and by movement it is converted to slate. New minerals are developed by recrystallization of the materials of the shales, and these along with any elongated minerals of the shale tend to be lined out in the direction of movement. Thus the minerals of the slate lie parallel, and this causes the slate to break into parallel layers. This property is slaty cleavage. Similarly a granite, under strong pressure and movement, will become a gneiss. Other igneous rocks also are changed to schists and gneisses.

Map Symbols.—For representing certain rocks on geological drawings, certain symbols are in general use (Fig. 26). The groups of beds make up the geological formations, and groups of formations make up the rock systems. From these is constructed a standard geologic column (page 14), which shows the relative ages of the rocks, the younger ones appearing above the older ones in the table. The table, read from the bottom up, may be regarded as the table of contents of the history of the earth. The beds are the leaves, the formations are the signatures, and the rock systems are the chapters of the history.

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CHAPTER III

THE ATMOSPHERE

The atmosphere is the blanket of air that covers the rocks and the waters of the earth. Its mass is less than a millionth part of that of the earth, but its activities and influences are far-reaching. Its presence is necessary to sustain the varied life of the earth, and it acts as a blanket to equalize the temperatures of the earth's surface. It serves as a medium for the transfer of water that is evaporated from lands and seas and that is precipitated as rain upon the earth, wearing away the rocks and transporting them to the seas. It is one of the agents of weathering, and it transports and deposits weathered material.

Composition of the Air.—Dry air consists of about 78 per cent nitrogen, 21 per cent oxygen, and 0.94 per cent argon by volume. Additional constituents include carbon dioxide, hydrogen, neon, helium, krypton, xenon, oxides of nitrogen, and ozone in minute amounts and, locally, certain volatile organic substances, sulphurous gases, chlorine, etc., from volcanoes and other sources. Water vapor also is an important part of the air, probably averaging about 1.2 per cent of the total volume. Its abundance varies according to the temperature; it forms about 2.63 per cent at the equator, 0.92 per cent at Lat. 50°N., and 0.22 per cent at Lat. 70°N. Fine earthy matter, smoke, soot, pollen, spores, bacteria, volcanic dust, meteoric dust, etc., may be spread as impurities through a considerable part of the atmosphere, sufficiently at times to darken the sun and reduce visibility.

Changes in Composition with Altitude.—The air extends to great elevations above the land. Mountain climbers have reached elevations of 27,000 feet on Mount Everest, and observers in balloons and airplanes have reached elevations of more than 13 miles. Balloons carrying meteorological instruments have reached elevations of approximately 20 miles; shooting stars, white hot from friction with the atmosphere, have been observed about 125 miles high; and auroral discharges are seen 375 miles above-ground (Fig. 27).

Because the air is heated chiefly at the bottom, the temperature of the air decreases upward about 1°F. for every 300 feet of difference of vertical elevation to altitudes of 6 to 8 miles, above which a zone of nearly constant temperature (about −67°F.) is reached. Differences in altitude account for the pronounced differences in temperature and corresponding differences in climate, in vegetation, and in habitability of places having the same latitude. A change in altitude of 1 mile in general is about equal to a change in latitude of 800 miles.

The dust and other earthy material in the air are confined essentially to the lower layers of the atmosphere. As one ascends into the air, he leaves the smoke and coarser dust behind. The water vapor becomes less and less, until at six or seven miles above sea level in the middle latitudes it is so cold that practically no moisture can remain in the air. Consequently no ordinary clouds exist. This altitude marks the lower limit of the *stratosphere*, a region of cold, clear, thin, dry air where there is a nearly constant temperature of about −67°F.

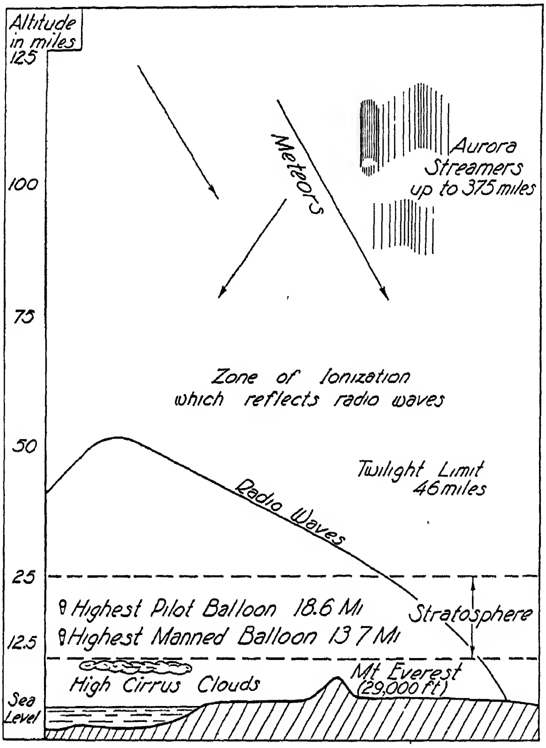

FIG. 27.—Diagram illustrating the height and character of the atmosphere. (*After Stetson, Humphreys, and others.*)

About one-half of the mass of the atmosphere occurs in the lower 18,000 feet. Thus at an elevation of about 3½ miles one is above more than half of the atmosphere. *Explorer* II, a United States Army balloon, reached an altitude of 72,395 feet, or 13.71 miles above sea level. It was above 96 per cent of the mass of the atmosphere.[1]

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[1] STEVENS, CAPT. ALBERT W., U.S.A., *Nat. Geog. Mag.*, vol. 69, p. 635, 1936.

At an elevation of about 25 miles the amount of ozone is much greater than near the earth's surface. Ozone is similar to oxygen, except that three atoms are linked together to form the molecule, instead of two as in ordinary oxygen. The ozone layer (Fig. 27) in the atmosphere absorbs or intercepts a high percentage of the ultraviolet rays coming to the earth from interstellar space. It has been estimated that if the ozone were decreased so as to allow even half of the ultraviolet rays to reach us, they would destroy our skins in a few minutes' exposure to the sun. However, if still more ozone were added to the atmosphere it would absorb so many of the ultraviolet rays that animal life on land would suffer because of a lack of the essential "sunshine" vitamin.

Sources of Heat.—The air is heated mainly by the sun. Additional but minor sources of heat include radiation from the interior of the earth and eruptions of steam and other hot gases from inside the earth. The amount of heat received by the earth from the sun is sufficient to melt a block of ice 1 mile square and nearly 100 feet thick every second or to melt a layer of ice about 15 feet thick over the entire earth in a year. The lower part of the atmosphere warms up more readily than the upper part on account of radiation from below and also on account of its greater density and its included water vapor. The most effective absorbents of heat in the atmosphere are water vapor and carbon dioxide. Barren rock surfaces absorb and later radiate heat more rapidly than areas covered with soil and vegetation or with snow and ice. It is noteworthy also that the land warms up and cools off faster than the sea, because rocks absorb and radiate heat more readily than water; they reflect less of the sunshine than water does; they are less deeply penetrated by solar radiation; and they are less affected by cooling due to evaporation. Air temperatures over both land and sea are modified, however, by ocean currents and by prevailing winds.

Weight of the Air.—The atmosphere weighs about 14.7 pounds per square inch at sea level, or enough to balance a column of water 32.8 feet high or a column of mercury in a barometer to a height of 29.92 inches, or 760 millimeters. Its weight rapidly decreases upward to one-half its total weight at an elevation of about 3.4 miles. Pressures in the lowest 6 miles are variable, but above that level they are nearly uniform at all times and places for any particular elevation. The variations of pressure are caused chiefly by heating and cooling and by differences in the amount of water vapor in the lower atmosphere. Increased temperatures cause the air to expand and make it lighter. Lowered temperatures and locally mechanical crowding of the air increase its density and pressure.

Distribution of Air Pressures.—Air pressures are measured by barometers which record the weight of the atmosphere above the place where the barometer is stationed. The aneroid barometer (Figs. 28, 29,) is simply a box of thin metal from which part of the air has been withdrawn and which responds to the pressure on the sides of the box. A mercury barometer is a tube filled with mercury and inverted over a bowl of it (Fig. 30). Increased pressure on the mercury in the bowl causes the mercury to rise higher in the tube, and the tube is graduated so that the height of the mercury is easily read. Readings over the earth's surface show that there is a belt of low air pressure along the equator and that the pressure increases to the north and south of the equator and is highest approximately along the parallel 30°N. and 30°S. (Fig. 31). From these belts, north and south, pressures decrease.

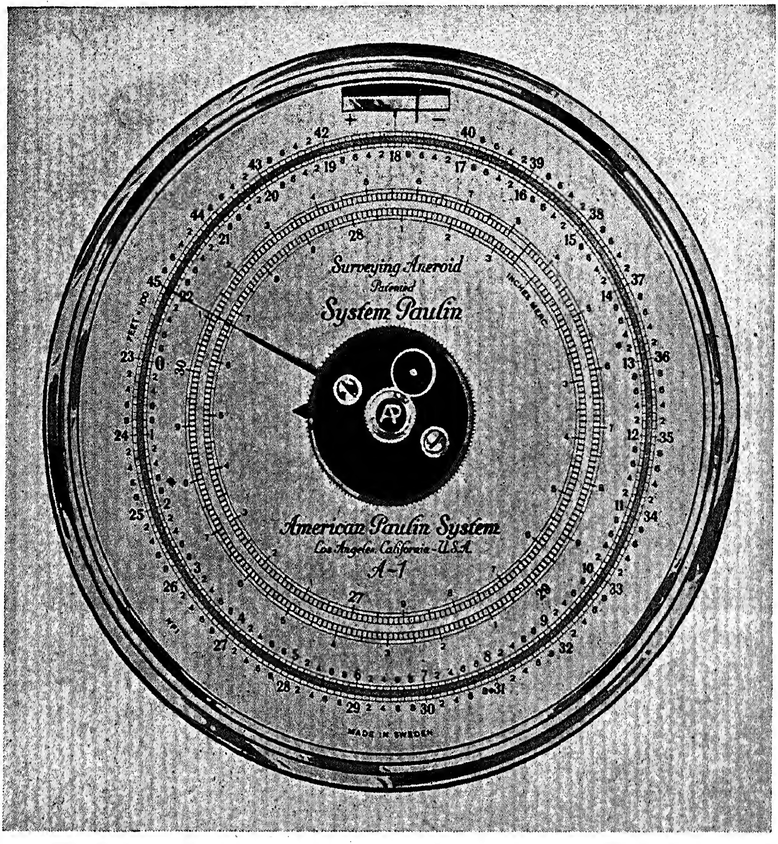

FIG. 28.—Dial of an aneroid barometer. The atmospheric pressure is indicated as inches of mercury on the inner dial and as elevation above sea level on the outer dial.

Movements of the Atmosphere.—Wind is air moving essentially along the surface of the earth. Movements of air up and down or far above the earth are "air currents." The cause of movements of the air s the heat of the sun.

If a gas jet is placed below the center of a broad glass dish or a pan filled with water, the water above the flame (Fig. 32) is heated, expands, and flows over to the sides of the dish, then down the sides of the dish and along the bottom toward the flame. This is shown if crystals that give off a colored solution are dropped at various places in the pan. The streams of colored water from all points move toward the flame, rise, and descend around the walls.[1] This experiment illustrates the movements of air.

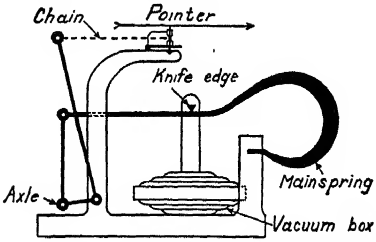

FIG. 29.—Diagram showing the construction of an aneroid barometer. (*Redrawn from Killen.*)

If the earth were at rest, there would be a constant movement up from the equator, then north toward the pole in the northern hemisphere, then a return to the south along the earth's surface. Such a movement may be compared to that of water in the pan (Fig. 32). There is a difference, however, for from the hot center of the pan outward to the cooler rim the areas become larger. They may be considered as wedges pointing to the hot column. In the earth the heated area is the equator, which is nearly 25,000 miles long; and coolest areas are near the poles, which are "points." The hot air overflowing the equatorial belt toward the poles would tend to bank up, because the poleward areas are so much smaller than the equatorial areas. In the latitudes of about 30° the banking up of the hot air from the equator results in high pressures, or "barometric highs" (see Fig. 31). There is so much air moving northward from the equatorial zone and there is so small an area for the air to cover in the poleward regions that the equatorial air cannot readily be accommodated. It piles up in the regions about 30° north and south, causing the high barometers in these belts. In these regions areas along the longitudinal circles of the earth are appreciably smaller than along the equator.

FIG. 30.—Diagram showing a mercury barometer. The tube is filled with mercury and is inverted over a cup with mercury. The mercury column falls until *AB* balances the pressure of the air at the place where the barometer is located.

The velocities of particles on several parallels of the earth is indicated by Fig. 33. Heated by the sun, the air rises at the equator and, in the

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[1] Small crystals of potassium permanganate will serve to show the directions of the currents. The pan should be a foot in diameter or more, and the bottom should be flat. In smaller pans currents are developed that move across the area above the flame.

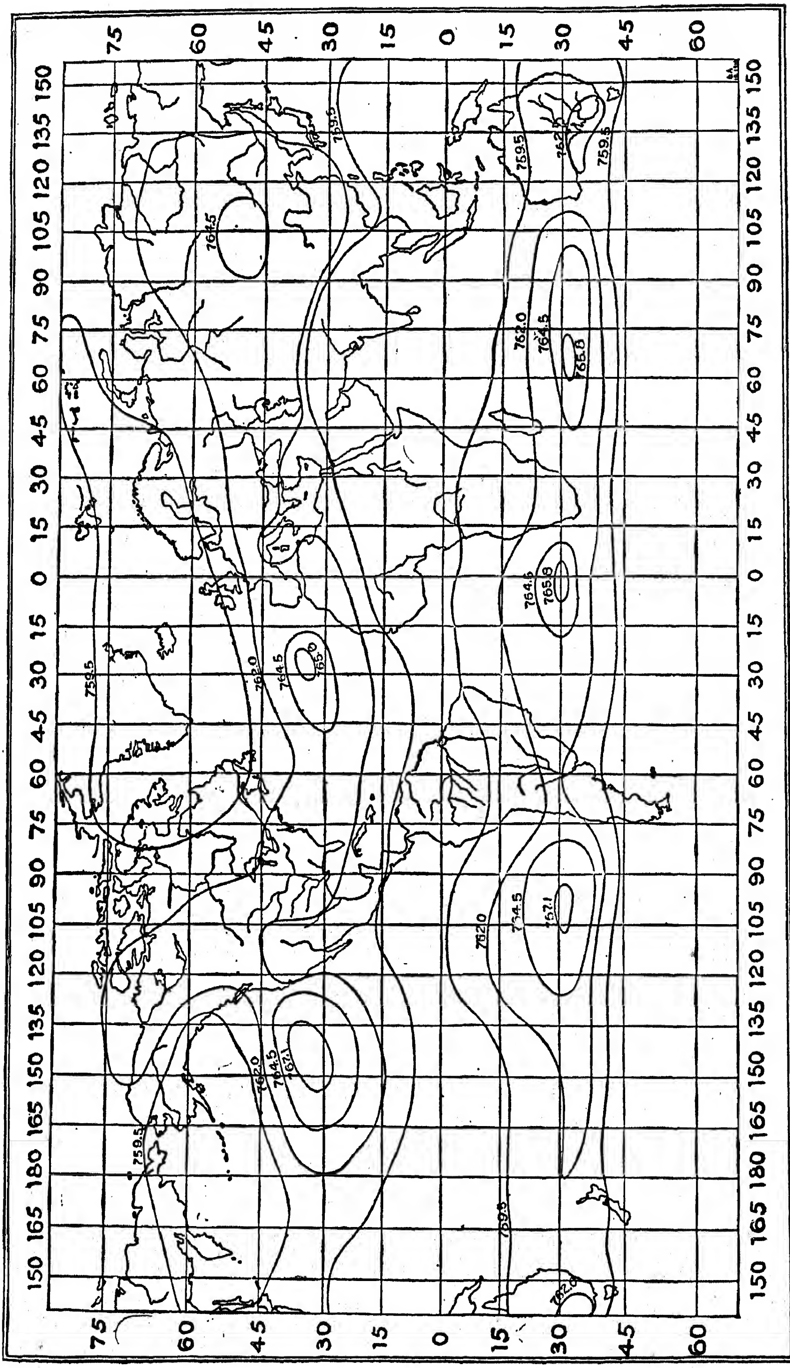

FIG. 31.—Map of world showing the average annual isobars. The belts of high barometer readings are about 32° N. and 32° S. latitude. The numbers are millimeters of mercury. (*After Buchan.*)

northern hemisphere, blows northeast (Fig. 34) as the antitrade wind (air current). At about 30°N, because it is getting cooler and because it is passing into a region where belts or areas along the parallels are smaller, some of the air is forced down. It becomes a westerly wind, the air moving to the east faster than the earth below it moves. The air soon slows down by friction with the earth and with other air, and some of the wind of the westerlies is drawn back to take the place of the air that flowed to the northeast from the equator as a southwest air current. This is the trade wind and blows from northeast to southwest (Fig. 34).

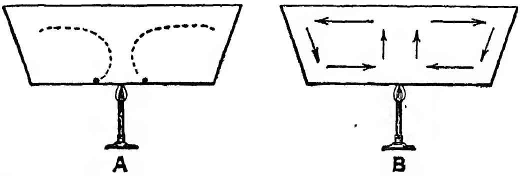

Fig. 32.—Diagrams showing the movement of water in a pan due to the heat from the gas jet below the pan. *A*, crystals of potassium permanganate dropped into pan show movement of water currents by coloring the water; *B*, arrows show the movement of water current in a heated pan.

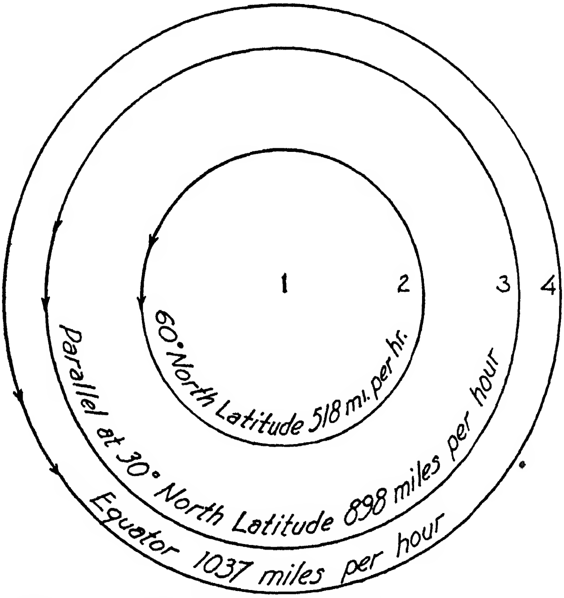

Fig. 33.—Diagram showing northern hemisphere of the earth as seen from a point directly above the north pole. The velocity of the earth's surface at three parallels is indicated in miles per hour.

Since the winds do not blow with nearly the great velocities required by the theory (*angular momentum conservation*) above stated, it is clear that there are also other active forces. The winds are slowed down to a few miles per hour by friction with the earth and with other bodies of air. The process outlined above operating in the belt from 30°S. to 30°N. latitude, is not the only cause of movement and probably is not the most important one, especially in the areas between 30°N. and 30°S. and the poles.

There is an important force known as the *horizontal deflective force.* This force may be illustrated by the pendulum. Suppose a long heavy pendulum is mounted so that there is no torsion, and is placed at the north pole so that its arc lies about perpendicular to the earth's axis. It would swing in the same arc, but its arc would seem to change position 15° per hour. This change results not from the change of direction of the pendulum but from the position of the earth with respect to the pendulum.

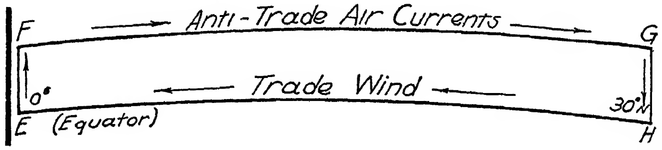

FIG. 34.—Air heated by the sun rises at the equator *E*, to *F*, where it overflows the air column *EF*, and moves northeast to about parallel 30°N. at *G*. There it banks up. A belt or an area covered at 30°N. is smaller than the area at the equator. The air is cooled and forced down and part of it becomes a westerly wind. It is moving east faster than the earth is moving at *H*. Later to replace the rising air column slowed down by friction it is drawn back from *H* to *E*. It is a trade wind blowing from northeast to southwest.

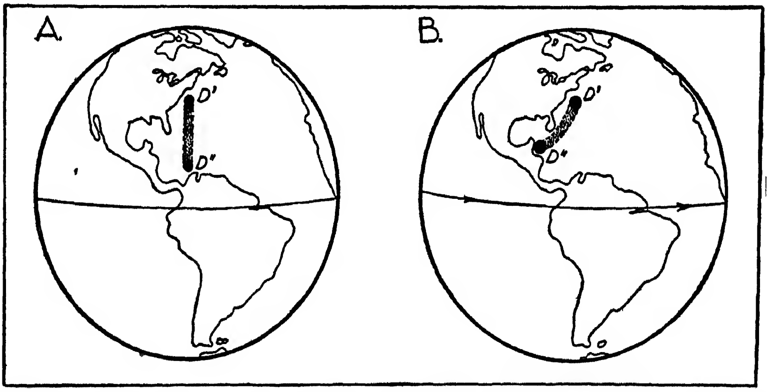

FIG. 35.—Diagram showing the influence of the rotation of a globe on the direction of travel of a drop of liquid on its surface. *A*, with the globe at rest, a drop of oil at *D'* moves downward to *D''*; *B*, on a globe rotated west to east, a drop of oil is deflected from *D'* to *D''*.

A pendulum mounted to swing across the equator at right angles swings in an arc parallel to the earth's axis and does not seem to change its position. There is no horizontal deflection because the arc of the pendulum and the axis of the earth are parallel. The amount of horizontal deflection gradually increases from the equator where it is 0° to the poles where it is 360° per day. This causes air moving south in the north latitude always to be deflected to the right or west, and in the south latitude to be deflected to the left.[1]

The deflection of the trade wind from south to southwest may be illustrated by rotating a small globe and dropping oil or water on it. The liquid moves down along a meridian of the globe when the globe is at rest but is deflected slowly to the right as seen from its starting point when the globe is rotated from west to east (Figs. 35, 36).

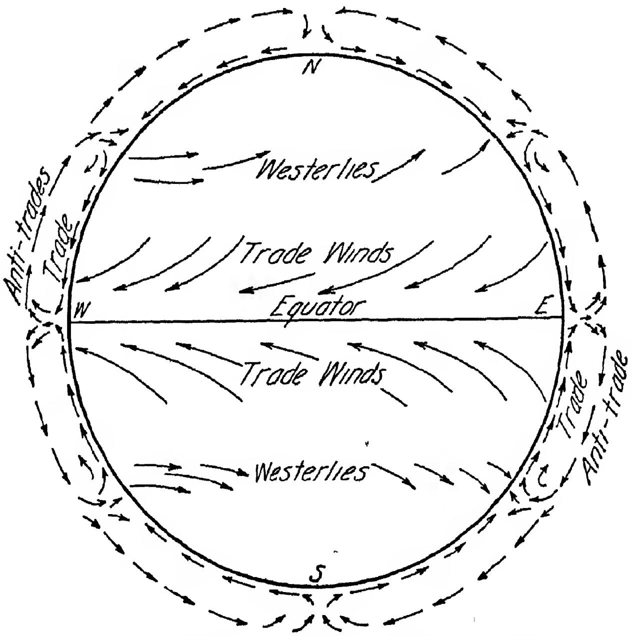

FIG. 36.—Ideal diagram showing major movements of winds and high air currents. (*Based on a drawing by Ferrel.*)

The configuration of the earth greatly modifies the air currents. They drag against the earth and against each other and are slowed down by the friction. The circulation of the air[1] is thus exceedingly complicated. It is modified greatly by local conditions, and it varies from day to day with temperature and precipitation and from other causes. The wind zones, trades, westerlies, and so forth, nevertheless are well defined and are of great value to those who navigate the seas.

Moisture of the Air.—The amount of water vapor in the air varies chiefly with the temperature. The warm air of the equatorial region may contain 3 to 4 per cent of water vapor, whereas that of the cooler middle latitudes may contain only 1 per cent or less. The average for all

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[1] A mathematical treatment by H. C. Willett is given in Physics of the Earth, III, Meteorology, pp. 133–233, National Research Council, 1931; and by W. J. Humphrey, Physics of the Air, pp. 94–205, McGraw-Hill Book Company, Inc., 1929.

latitudes, according to Humphreys, is about 1.2 per cent. The upper atmosphere, or stratosphere, on account of its frigid temperature, contains practically no water vapor.

Condensation and Precipitation.—Condensation of moisture is caused by cooling and results in precipitation of dew, rain, snow, hail, and sleet. Cooling may be caused by radiation, by contact with cold surfaces, by mixing of masses of air of different temperatures, by expansion due to movement of the air to places of lower pressure, etc. Condensation to droplets of fog or cloud is facilitated by the presence of dust particles.

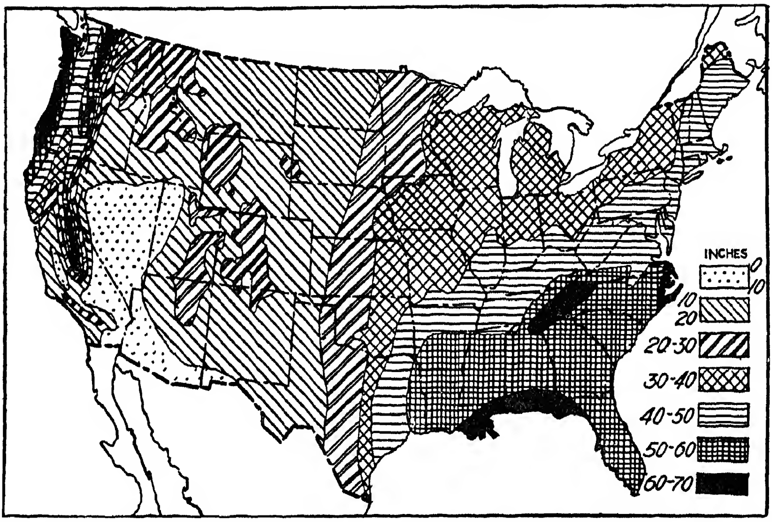

FIG. 37.—Outline map of the United States showing mean annual rainfall. (*After Fuller, U. S. Geol. Survey.*)

Rainfall.—The average rainfall of the earth is about 30 inches a year,[1] but about 20 per cent of the land receives less than 10 inches a year, and about 2 per cent receives more than 100 inches a year (Fig. 37). Rainfall on land is influenced mainly by (1) latitude; (2) nearness to the sea; (3) topography of the land, especially the presence of mountain ranges which intercept moisture-laden winds; (4) prevailing winds; (5) seasons; and (6) the frequency of cyclonic storms, hurricanes, and typhoons. Rainfall is heaviest in the tropical belt of calm where warm, moist air rises abundantly and where daily showers are the rule and on the windward sides of continents and mountain ranges. By contrast, many of

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[1] The precipitation on the earth's surface has been estimated very closely by Brückmann, who states that the annual precipitation on the entire earth's surface is 74.3 centimeters, or 29.5 inches. Of this amount 30 per cent falls on the land, which amounts to 26,679 cubic miles per year. Several other estimates state that the average annual precipitation is about 36 inches. Walter Brückmann, Leitfaden der Wetterkunde, pp. 1–284, Vieweg und Sohn, 1927.

the great deserts of the earth lie along the high-pressure belts or in the lee of mountains.

GRADATIONAL WORK OF THE ATMOSPHERE

Weathering.—The earth's crust is constantly changing. Even the rocks of the earth, which to a casual observer appear immutable, are subject to alterations which finally reduce them to different states, different physical forms, and different compositions. The processes by which rocks are so altered at the earth's surface are comprised under the term "weathering." This attack on rocks is produced largely by the access to the rocks of air and water, the weathering "elements." Thus weathering is related to the surface of the lithosphere, where rocks, air, and water come together. The weathering of rocks may be compared to the decay of a building. A house constructed of the strongest and most resistant stone in a few centuries will decay and fall in ruins unless it is continually repaired. Monuments, gravestones, roofing slates, roads, foundations, concrete buildings, steel bridges, and all other structures are subject to weathering in the same manner. Man continually is in contest with the weather. He chooses the resistant materials for buildings, and he paints exposed surfaces to delay disintegration, but ultimately all structures are destroyed by weathering and must be rebuilt.

FIG. 38.—Spheroidal weathering of basic igneous rock, north shore of Lake Superior. (*Photograph by F. F. Grout.*)

Water soaks into the rocks, dissolves and alters minerals, expands by freezing, and enlarges joints and fractures. The process is begun on cracks or fractures and ultimately affects the entire rock. The penetration of the elements along openings is illustrated by Fig. 38. The rounded spheroidal particles between the openings are last affected, but ultimately these also succumb to weathering. Where rock surfaces present materials of different hardness, the soft materials are weathered most readily and become cellular or pitted (Fig. 39).

Weathering is partly physical and partly chemical; the two phases are disintegration and decomposition. By disintegration is meant the physical disruption of rocks to form particles of smaller size without change in composition. The particles are of the same material, and the minerals are fresh. Decomposition, on the other hand, is chemical decay by which the rocks are broken down by chemical alteration of the minerals. Normally both phases of weathering are going on at the same time, and physical disruption of rocks facilitates the access of the chemical materials involved in rock decay. In relation to erosion, weathering is but the first step—the preparation of the materials for removal.

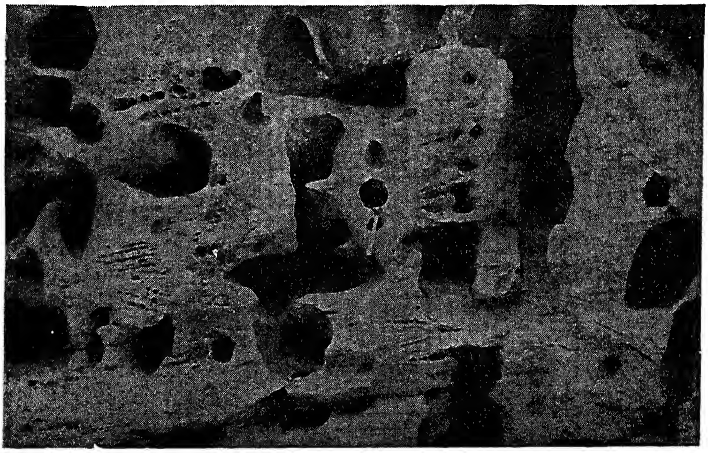

Fig. 39.—Solution pits due to weathering of sandstone. The softer parts of the sandstone have been washed out or blown away, leaving pits and lines of pits along the softer beds. (*Photograph by C. E. Erdmann, U. S. Geol. Survey.*)

Disintegration. *Temperature Changes.*—Just as steel bridges and concrete pavements are subject to expansion when heated and to contraction when cooled, so rocks are affected by the alternate heating and cooling, due to daily and seasonal changes in temperature. Repeated expansion and contraction thus tend to develop cracks in rocks. Water, entering these and other cracks, freezes and thaws. On mountain tops in the daytime the air may reach a temperature of 120°F. or more, and the rocks may become distinctly warm or even hot to the touch of a hand, but at night the temperature of the air drops below freezing. Under such conditions particularly, disintegration is effective. Hence, most high mountain peaks which are unprotected by snow are covered with fields of boulders. In time, the mountain tops may be reduced to dome-like forms mantled over with a sheet of accumulated débris.

Accumulations of rock fragments dislodged from cliffs by weathering and deposited below by gravity are *talus* (Fig. 40). The slope of the talus is approximately the angle of rest of the material, generally about 25 to 35 degrees from the horizontal, according to the size and angularity of the fragments, the amount of subsequent rainwash, and other conditions. Smooth rocks exposed to weathering commonly spall off in thin slabs, sheets, or scales concentric with the surface. On account of the low conductivity of rocks, which tends to confine the heating and cooling to the outer part of the mass, the outer part on expanding and contracting pulls away from the inner part until finally it falls off and exposes a fresh surface to the attack. Increase in volume on account of chemical changes, particularly that which is due to the addition of water or hydration, assists separation. The process of shelling or scaling off in such slabs, or leaves, is exfoliation (Fig. 41 *A*, *B*). This process leaves smooth, rounded surfaces on most rocks regardless of their composition. Striking rock domes, such as Stone Mountain in Georgia, or the great domeshaped outcrops in central Texas, have been formed by exfoliation.

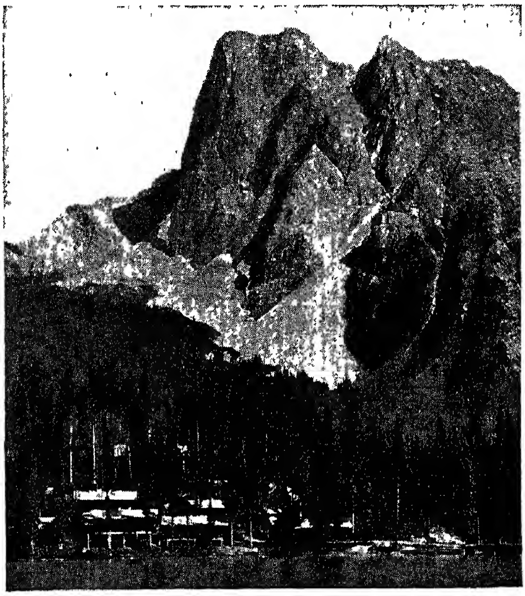

FIG. 40.—Talus slopes at base of the mountains near Emerald Lake, Yoho Park, British Columbia. By weathering, pieces of the cliffs are broken off and roll down to the base of the cliff, forming the talus pile. (*Courtesy Department of the Interior, Canada.*)

As the process of disintegration continues, the exfoliated slabs and chips of rock are subjected to other stresses. This is true especially if the rock is composed of two or more minerals. In granite, for example, various minerals react differently. The dark minerals such as hornblende and biotite absorb heat more readily and also give it up more quickly than the lighter colored feldspar and quartz. Furthermore each mineral has its own coefficient of expansion and contraction. Consequently, in an intergrowth of minerals such as occurs in a granite, stresses are set up which tend to separate the minerals. Finally the mineral grains fall apart and produce a sand of loose minerals. This process of *granular disintegration* is in operation over large areas where coarse-grained rocks are exposed to differential heating. Many of the slopes of Pikes Peak, Colorado, are covered with such products. The finer grains are transported by wind or water, and the coarser fragments remain until they are disintegrated more completely. At Dogtown Common, Cape Ann, Massachusetts, large granite boulders have crumbled into heaps of crystalline sand; and at Medford, Massachusetts, the exfoliated slabs of a diabase dike are disintegrated and decomposed into a soft granular mass of dark minerals.

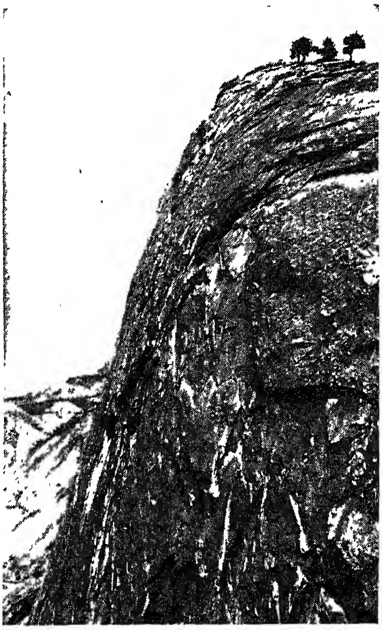

FIG. 41*A*.—Exfoliating granite, Yosemite National Park, California. This rock mass once had angular edges and irregular points which by exfoliation now are largely transformed into smooth curves. (*After Matthes, U. S. Geol. Survey.*)

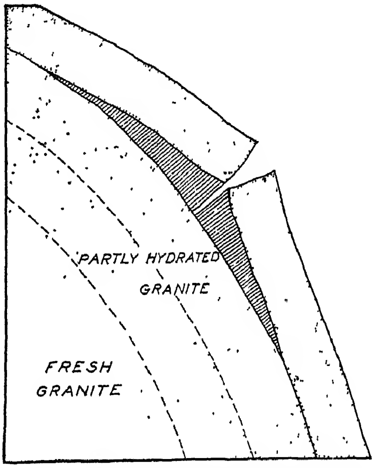

FIG. 41*B*.—Diagrammatic cross section of part of an exfoliating boulder of granite. It shows, lower left corner, the fresh or unaltered granite. Expansive force due to hydration has caused the outermost shell to crack and separate from the inner portion. (*After Blackwelder.*)

Freezing of Water.—The freezing of water in pores and cracks in rocks tends to disintegrate the rocks, for water in freezing expands about one-eleventh of its volume and exerts great pressure. Although the pressure is much less than the crushing strength of most hard rocks,[1]

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[1] The crushing strengths of rocks used for construction purposes generally range as follows: sandstones, 4,000 to 16,000 pounds; limestones, 6,000 to 12,000 pounds; granites, 15,000 to 30,000 pounds per square inch.

it is sufficient to disrupt soft rocks or rocks weakened by cracks or by partial decay. Highly porous rocks such as sandstones, whose pore space commonly ranges from 10 to 30 per cent, are disrupted in this way, and so are jointed rocks whose cracks become the loci of ice wedges. In the same manner the soils on fall-plowed lands in the north central part of the United States are made light, fluffy, and easily worked by repeated freezing and thawing during the winter and early spring. Frost heaving also brings boulders to the surface, and it pushes up posts, stakes, foundations, and other structures.

FIG. 42.—Rock being split by the growth of the roots of a tree. (*Photograph by Bastin, U. S. Geol. Survey.*)

By the combined action of frost and differential heating, unusual erosion forms may be produced. "The Old Man of the Mountain" in the White Mountains of New Hampshire is an outstanding example.

Plants and Animals.—Plants and animals play a prominent part in weathering (Fig. 42). Roots grow into cracks and crevices and push the fragments up and apart often as much as several feet. The overturning of well-rooted trees by the wind fractures the rocks and exposes them to destruction. The burrowing of animals such as earthworms,

FIG. 43.—Rock sculpturing produced by differential weathering. The group of ghost-like figures has been carved out of massive rock by the work of the atmosphere. Wheeler National Monument, Colorado. (*Photograph by C. Beery.*)

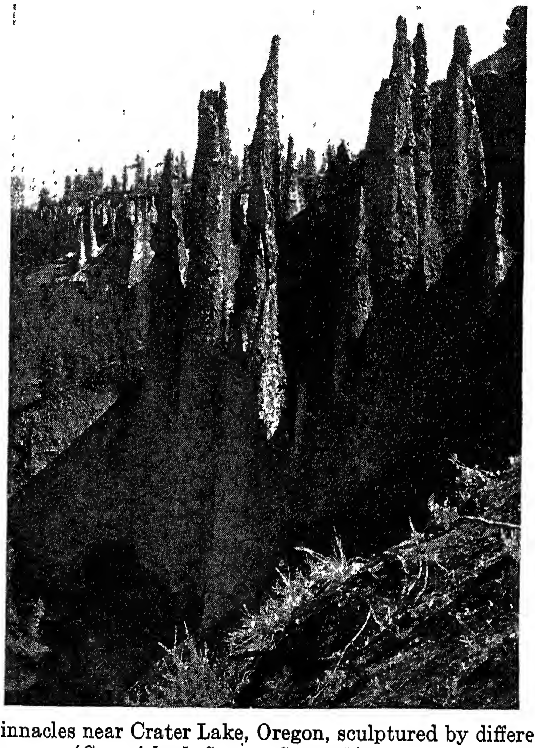

FIG. 44.—Rock pinnacles near Crater Lake, Oregon, sculptured by differential weathering. (*Copyrighted, Sawyer Scenic Photos, Inc.*)

ants, and rodents and the tramping of animals, especially hoofed mammals, also contribute to the disintegration of rocks. Man likewise does his part in excavating road cuts and tunnels, in quarrying, in mining, and in the cultivation of the land. The "breaking" of the sod on the prairies, the clearing of brush and timber, and the destruction of forests by lumbering and by fire have upset the previous balance between weathering and erosion and indirectly have permitted rapid erosion and renewed weathering over large areas.

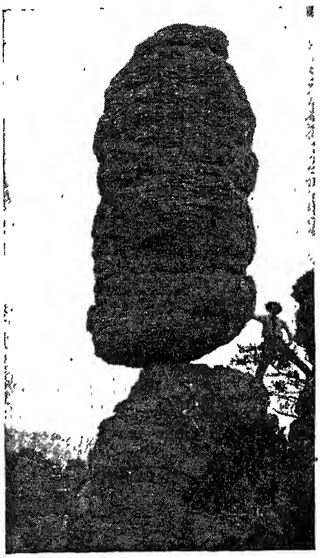

Fig. 45.—A residual balanced boulder near Douglas, Arizona, formed by the weathering and removal of the rock that once surrounded it. (*Courtesy Irwin Studio.*)

Decomposition. *Oxidation.*—Rocks decompose or decay by chemical alteration of their component minerals by oxidation, hydration, carbonation, and solution (Figs. 43, 44, 45). In the process of oxidation, oxygen is added to the rocks, especially to the iron compounds. The air contains about 21 per cent oxygen by volume. The oxidation of rocks by air is aided by the presence of moisture; without water, oxidation is generally slow. Air and water break down the ferrous silicates such as pyroxenes, amphiboles, and olivine, and the ferrous iron is converted to ferric oxide (hematite) or to hydroxides (goethite, limonite) with accompanying color changes from green or black to red, yellow, or brown. Hence many soils in warm, moist climates are colored red, yellow, or brown. Locally deoxidation or reduction by organic matter may occur, and near the roots of trees and under peat bogs the bright colors may be changed to somber ones. The oxidation of pyrite, which is composed of iron and sulphur,[1] leads to the formation of sulphuric acid which attacks the rocks and develops solution pits and accompanying stains and discolorations, so that even small amounts of iron sulphides may be injurious in building stones. The change of color of certain roofing slates from green or gray to brown is caused largely by rusting of the component iron compounds, as in normal weathering. In the oxidation of pyrite the sulphur as well as the iron is oxidized.

[1] The reaction is written

$$\underset{\text{pyrite}}{FeS_2} + \underset{\text{water}}{H_2O} + \underset{\text{oxygen}}{7O} = \underset{\substack{\text{ferrous}\\\text{sulphate}}}{FeSO_4} + \underset{\substack{\text{sulphuric}\\\text{acid}}}{H_2SO_4}$$

Hydration.—Hydration involves the chemical addition of water to the minerals of a rock to form new minerals, chiefly hydrous silicates and hydrous oxides. Thus orthoclase, a mineral abundant in granite, is broken up and converted largely to kaolin, the principal mineral in common clay. The potassium and excess silica are released at the same time.[1] Plagioclase feldspars are decomposed in the same way, and most of the alumina likewise is used in forming kaolin. Other hydrous silicates formed by hydration of the primary silicates include chlorite, serpentine, talc, zeolites, and many other minerals.

Certain aluminous minerals like nepheline[2] contain less silica than the feldspars and by prolonged weathering yield bauxite, diaspore, and gibbsite. Gypsum, calcium sulphate, which contains water, commonly is formed by the hydration of anhydrite.[3]

Certain minerals are oxidized and hydrated at the same time. Thus pyrite may be changed to limonite by the action of water and air.[4]

Carbonation.—Another process of decomposition is carbonation, by which carbon dioxide, CO_2, is added to certain bases, particularly to oxides of calcium, magnesium, sodium, and potassium, to form carbonates or bicarbonates of these metals. All ordinary waters contain dissolved carbon dioxide, which is derived from the atmosphere. Carbonated water dissolves many substances more readily than pure water, and it is consequently an active agent of weathering.

It is noteworthy that the carbonated waters, although only feebly charged, are very abundant and that the carbonates of the alkalies and alkaline earths are soluble in such waters, so that solution of these materials goes on together with carbonation and plays an important part in the decomposition of rocks by removing certain constituents. In addition to calcium, magnesium, sodium, and potassium, even the less soluble silica, alumina, and iron are taken away in part. Dissolved sulphates and chlorides are less abundant than the bicarbonates. Certain rocks particularly are subject to solution—notably limestone and marble, both composed chiefly of calcium carbonate, which is soluble in solutions

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[1]
$$\underset{\text{orthoclase}}{2KAlSi_3O_8} + \underset{\text{water}}{2H_2O} + \underset{\substack{\text{carbon}\\\text{dioxide}}}{CO_2} = \underset{\text{kaolin}}{Al_2Si_2O_5(OH)_4} + \underset{\text{silica}}{4SiO_2} + \underset{\substack{\text{potassium}\\\text{carbonate}}}{K_2CO_3}$$

$$\underset{\text{albite}}{2NaAlSi_3O_8} + \underset{\text{water}}{2H_2O} + \underset{\substack{\text{carbon}\\\text{dioxide}}}{CO_2} = \underset{\text{kaolin}}{Al_2Si_2O_5(OH)_4} + \underset{\text{silica}}{4SiO_2} + \underset{\substack{\text{sodium}\\\text{carbonate}}}{Na_2CO_3}$$

[2] Nepheline ($NaAlSiO_4$), yields diaspore ($Al_2O_3.H_2O$), gibbsite ($Al(OH)_3$), and bauxite, a mineral with Al_2O_3 containing more water than diaspore and less water than gibbsite. Bauxite is the chief ore of aluminum. It is a mixture of several hydrated aluminum oxides.

[3]
$$\underset{\text{anhydrite}}{CaSO_4} + \underset{\text{water}}{2H_2O} = \underset{\text{gypsum}}{CaSO_4.2H_2O}$$

[4] This reaction is somewhat complicated. The product, limonite, is $Fe_2O_3.H_2O$.

that carry carbon dioxide. Gypsum and rock salt are very readily soluble in water.[1] Exposed surfaces of limestone and gypsum generally become etched or pitted by solution. Rocks that are made up of two kinds of material, one readily dissolved and another less readily dissolved, develop pitted surfaces in which the less soluble material stands out in relief.

Work of Vegetation.—Vegetation assists decomposition. Lichens, which are among the first plants to grow on freshly exposed rocks, take certain chemical elements from the rocks, and the roots of other plants take up additional inorganic matter. Furthermore, the decay of the organic matter itself releases certain organic acids which increase the solvent power of the natural waters. The solubility of silica, alumina, and iron, for example, is much greater in the presence of these organic acids. The chemical activity of the small but abundant and ever present bacteria which produce ammonia, nitric acid, carbon dioxide, and other active chemical compounds is another factor in the alteration of rocks and in the formation of soils. Indirectly vegetation serves to retain moisture and to delay erosion and hence to prolong chemical weathering.

Depth of Decomposition.—Rock decomposition may proceed to great depths. Granitic rocks in the District of Columbia are decayed to a depth of 80 feet, and near Atlanta, Georgia, similar rock is decayed to approximately 100 feet. In northwestern Georgia the depth of decay of limestones is nearly 200 feet, and in Brazil shales are decayed to a depth of 400 feet.

The Influence of Climate on Weathering.—The nature and extent of weathering is controlled largely by climatic conditions. There are many factors that bring about climatic variations, but four distinct types of climate may be recognized: (1) the hot and moist climate of the equatorial belt, (2) the hot and dry desert climate, (3) the cool and moist climate of the temperate zones, and (4) the cold and dry arctic regions. In each of these regions, rock weathering is going on continually, but each region presents peculiarities of its own.

In a moist, warm climate, rock decay is rapid. In a dry climate it proceeds more slowly. The effect of climate is well illustrated by the Egyptian obelisk that was presented to New York City. It had stood without apparent injury for many centuries in the mild dry climate of northern Egypt but began to disintegrate soon after its removal to Central Park so that special protection had to be given to it.

In the equatorial regions where the rainfall is heavy and the temperature is high, chemical processes are active, and the influences of organic

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[1] Commercial beds of gypsum, $CaSO_4.2H_2O$, and rock salt, $NaCl$, commonly are protected from solution in humid regions by a cover of relatively impervious beds of clay or shale.

agencies are pronounced. The chemical reactions are more rapid than in cooler latitudes, and consequently the decomposition of silicates is more complete and much silica is removed in solution. The end product of such weathering is *laterite*, which consists largely of red hydrated oxides of aluminum and iron. This lateritic residue takes the place to a large extent of the clayey mantle rock of the higher latitudes.

In desert regions peculiar conditions prevail, and therefore the character of weathering differs from that which is found elsewhere. Rainfall is sparse, so that solution by downward-percolating water is of minor importance. Some water, however, is retained by capillary action. Since the air is dry and the sun is hot, the capillary water is drawn toward the surface where evaporation concentrates the salts that are in solution. These warm concentrated solutions react with the constituents of the rocks and tend to decompose them. The crystallization of new compounds takes place between the mineral grains, and these may cause splitting of solid rocks, in much the same manner as the freezing of water. Since waters are constantly rising toward the surface by capillary action and depositing their dissolved contents, the surface materials become cemented, forming "hardpan" and irregular concretionary masses in the mantle rock. Because of excessive evaporation, the soluble salts of sodium, calcium, and magnesium commonly occur as efflorescence on the surface. This is especially true in depressions and over flat areas, where they remain because of the inadequacy of the rainfall to wash them out. In many desert regions a brown or black shiny crust on the rocks is known as "desert varnish." It consists mainly of oxides of iron and manganese. Desert varnish has been thought generally to be the result of deposition of mineral matter from evaporated capillary water. Recent studies indicate, however, that in some instances the growth of lichens may be an important contributing factor.

In temperate regions there are marked seasonal variations in climate, and consequently the type of weathering is to a certain extent a combination of all others. In winter, frost action is dominant; whereas in summer, spring, and autumn, percolating waters play a more important part. The elevation above sea level also is an important factor, especially if the mountains extend above the timber line. Low-altitude temperate regions generally are not subject to extreme and sudden changes of temperature. In general, solution and chemical decomposition are the dominant types of weathering.

In the subpolar regions, where a large part of the surface is covered with snow during most of the year, the underlying rocks are saturated with thaw water which is repeatedly frozen and thawed. By far the most important weathering agent in such an environment is the expansion of water when it freezes. This process shatters the rocks and leads to the accumulation of a mantle of angular fragments. Essentially similar conditions prevail in the high, snowcapped mountains of the temperate and equatorial regions.

Formation of Soils.—Most of the land areas of the earth are covered by thin layers of decayed or disintegrated rock directly due to weathering, to which the term mantle rock commonly is applied. Soil proper is the thin upper portion of this mantle which is decomposed and altered sufficiently to support plant life. It varies in depth from several inches to three or four feet or more, and generally it is colored dark by decayed organic matter. Some soils rest upon the bedrock from which they are derived (Figs. 46, 47). Such soils are residual. Transported soils are those which are derived from mantle rocks that have been carried to their present positions from their places of origin (Figs. 48, 49).

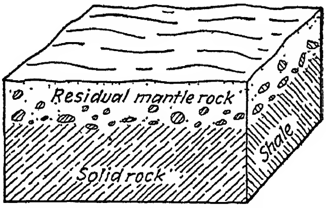

FIG. 46.—A residual mantle rock formed by weathering of the underlying shale. Fragments of the parent shale are found in the mantle rock and generally lie with the same attitude as the parent rock.

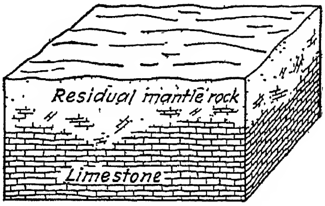

FIG. 47.—Residual mantle rock formed by weathering of underlying limestone. Fragments are found in the mantle rock, and many of them lie horizontal like the underlying limestone.

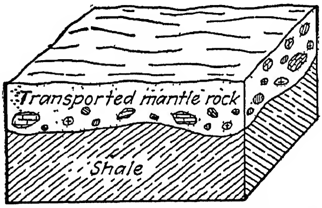

FIG. 48.—Transported mantle rock. The material of the mantle rock is different from the tilted solid rock and contains different kinds of fragments. These fragments lie in different positions.

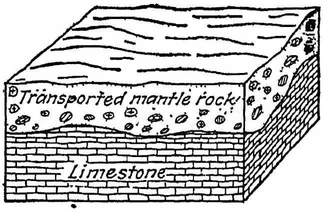

FIG. 49.—Transported mantle rock. The material of the mantle rock is different from the flat-lying solid rock and contains different kinds of fragments, which lie in many different positions.

Residual soils show a gradual transition downward into subsoil, and this in turn grades imperceptibly into rotted rock, which crumbles readily when exposed at the surface. The thickness of the subsoil varies greatly, for the depth to which rock is decayed is greater at some places than at others. Furthermore, owing to gravity the products of disintegration do not accumulate on steep slopes. Where the slopes are considerable, solid rock crops out at the surface, not because no mantle rock is formed but because the loose material is removed by rainwash, wind action, or other agents of erosion as rapidly as it is formed. Figure 50 shows solid rock grading upward through subsoil to soil. On the steep slopes at the right no soil has accumulated, even though the rock is disintegrating.

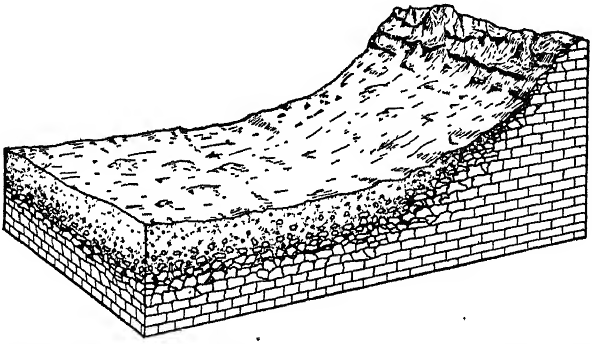

FIG. 50.—Diagram illustrating the formation of soil on the side of a hill. The solid rock in the valley grades upward to partly decayed rock, which in turn grades into soil. On the steep slope no soil has accumulated, because it is carried away.

Because rocks differ in mineral composition, the soils derived from them differ also both in texture and in composition. Some are coarse and sandy, others fine and clayey; some are sterile, others fertile. When granite is weathered under humid conditions, a residual soil is formed containing grains of quartz sand in a matrix of clay. Such a residue is referred to as a loamy soil. The decomposition of limestone leaves a thin layer of a finely textured clayey soil. Limestone consists mainly of calcium carbonate or the mineral calcite. This mineral is dissolved by the agents of weathering, and therefore the amount of residue is not great and consists mainly of fine-grained impurities that were in the limestone. Some limestones contain phosphatic shells which liberate phosphates that remain in and enrich the residual soil as the rock is disintegrated. The blue-grass basin of Kentucky is an example (Fig. 51).

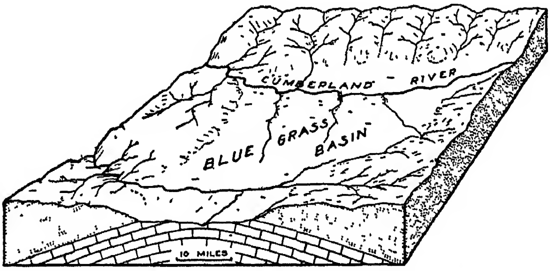

FIG. 51.—Diagram showing the relation of the fertile "blue-grass basin" soil of Kentucky to the underlying phosphate limestone. Poorer soils derived from sandstone and shale occur in the highlands surrounding the basin.

Transported soils are made up largely of material that is not weathered or that is only partly weathered. They owe their present positions to some agent of transportation, such as running water, wind, moving ice sheets, or gravity. Since these agents accomplish different degrees of sorting, transported soils vary in texture from fine silts to coarse gravel. They vary also in chemical composition, and certain transported glacial soils are very fertile, since they are made up largely of ground-up rocks that have not been leached by water of certain valuable mineral foods that plants require.

Replenishment of Soils.—Year by year rocks disintegrate to form the soils. Year by year the soils are carried away by the brooks to the creeks and thence by the rivers to the sea. The farmer may observe his rich, dark soil being gradually washed away, and often he provides small dams and earthworks to prevent wash. The soil commonly is deposited over the flood plains of the rivers and over deltas and flats, and these are noted for their fertility. When the top of the soil is washed away, however, new rock matter is exposed to weathering, and this provides additional mineral substances which are necessary for plant life. Plants utilize a large number of materials for their growth. These are derived from the atmosphere and from the soil. They include carbon, hydrogen, oxygen, nitrogen, phosphorus, sulphur, potassium, calcium, and other substances. The carbon, hydrogen, and oxygen are obtained from the air and water, and nitrogen may be developed in the soil from the atmosphere by a proper rotation of crops. The clovers, beans, peas, and similar crops add nitrogen compounds to the soil. Iron, calcium, magnesium, and certain other substances are desirable in the soil, but these are generally found in soils, and in many soils they are present in abundance. On the other hand, phosphorus, potassium, and nitrogen compounds often are added as commercial fertilizers. In general, phosphorus is present in igneous rocks as the mineral apatite (calcium phosphate) and in sedimentary rocks as a nearly similar substance. Potassium is present in orthoclase and in many shales and other sedimentary rocks. When soils are cultivated continually for many years, however, and particularly when grain is harvested and removed from the land, there is a steady decrease in fertility. When soil has been exposed to weathering for ages with very little erosion and very little removal of rock matter to expose new minerals to decay, the mineral fertilizers are essential.

Mineral Fertilizers.—The chief mineral fertilizers are nitrates, phosphates, potash salts, and calcium salts. All of these are added to certain soils in large amounts. Nitrates are obtained from the desert regions of Chile, where they have formed by the drying up of waters containing sodium nitrates and other salts. In recent years much nitrate has been made artificially from nitrogen of the air. Potash salts have been imported largely from Germany where they are found with salt and gypsum in beds that have resulted from the drying up of ancient seas. Recently large amounts of potash salts have been found in West Texas and in New Mexico. Calcium phosphate, as stated, is present in nearly all igneous rocks as apatite, and that is its chief source in the natural soils, but the calcium phosphate of commerce is obtained chiefly from "rock phosphate" which is found in sedimentary beds. The phosphate is made more readily available by treatment with sulphuric acid. Calcium is added to the soil as calcium sulphate (gypsum) and calcium carbonate (powdered limestone). Often these substances are mixed together as a fine powder and planted with the grain so that only small amounts are required annually. In the United States the mineral fertilizers are used chiefly by cotton planters and fruit and vegetable growers.

WORK OF WINDS

Nature of Work.—The wind derives its energy from the heat of the sun and expends a part of that energy in moving dust and sand about over the face of the earth. Since these materials are land derived, such wind work is more important over the land than over the sea; nevertheless wind-borne dust and silt may be carried far out to sea and ultimately may be dropped into ocean waters.

As compared with streams of water, air currents lack the concentrated thrust and the steadiness of great rivers, so that the work of wind on the whole is less important than the work of streams. Its great field of activity, however, is in arid regions, which are more or less devoid of vegetation and are plentifully supplied with fine rock waste, and where stream work is at its minimum. For this reason the wind tends to compensate for the reduction of the work of streams in arid regions and to supplement their activity. Wind work, however, is not confined to arid regions, for even the most humid regions have dry seasons favorable to wind activity, and wind-borne dust may be carried to them notwithstanding their prevailingly humid climate.

Methods of Wind Erosion.—In loose, dry materials the impact of the wind itself is sufficient to remove vast quantities of earthy matter by the process of *deflation* (Latin, *deflare,* to blow away). Eddies, whirlwinds, and updrafts help the wind to lift and remove its load in this way. Thus many exposed uplands are swept free of loose material which is carried away as fast as weathering produces it. The so-called stony deserts are produced by the removal by wind of all fine products of disintegration, leaving only coarser *lag gravels* and boulders strewn over a rocky surface. Contrary to popular conception, only small parts of deserts are covered with drifting sands. The Sahara desert covers 3,500,000 square miles.

Less than one-half million square miles is dune covered; the remainder has a boulder-strewn rocky floor. A more spectacular but perhaps less important method of wind erosion is that of *corrasion* and *abrasion* (Latin, *abradere*, to scrape off or rub away), *i.e.*, the use of a natural sand blast whereby the wind employs its load of sand as a cutting tool in the same manner as man cleans stone or brick buildings or cuts figures on stone with the aid of air-driven sand as an abrasive. By means of the sand swept along by the wind, rock surfaces are scoured and grooved. In desert regions such grooves are very conspicuous, and where the winds have slight variation of direction the grooves show a parallel alignment and their sides are often strongly fluted. Many odd-shaped land forms are developed. These include undercut hills, with accompanying broad shallow caves, mushroom rocks, table rocks, pedestals, and similar sculpturings (Fig. 52). Deep hollows may be developed in loose or easily eroded material. In the Gobi desert these hollows range from about 300 yards to 30 miles or more in length and from 50 to 400 feet in depth. The depths are limited by the position of the regional ground-water level.

FIG. 52.—Pedestal rock sculptured by wind erosion, Montana. (*Photograph by Erdmann, U. S. Geol. Survey.*)

The sand grains used as tools in the natural sand blast are themselves subject to wear, so that they are chipped, pitted, and generally reduced in size. Examined with a lens, the battered grains may show a "frosted" surface like that shown by ground glass and concentric cracks like the familiar "moon" texture in marbles.

Special Effects of Wind Erosion.—Occasionally one may see fields of young wheat or corn ruined by the removal of soil by deflation to such an extent that the tender roots are exposed to a withering sun. Fallow land in dry-farming areas especially is subject to damage in this way. An inch or two of soil thus may be taken away by the wind in the course of a few days. Many examples of wind abrasion, which have been recorded by man, may be given. These include the "frosting" and ultimate destruction of glass windows exposed along sandy seashores, the cutting down of telephone poles just above the ground in deserts, the marring of the Sphinx and of certain pyramids in Egypt, the undercutting of stone foundations of buildings, and many others. From these examples it is only a short step to a similar interpretation of many oddities of windswept arid regions. These include beveled stones—*einkanter* or *dreikanter* (meaning one- or three-edged) (Fig. 274).

Einkanter apparently are formed by the cutting of pebbles under conditions of a constant direction of the wind; dreikanter, shaped like Brazil nuts, suggest that the pebbles were overturned, perhaps as a result of undermining, so that several facets are developed in succession. The polish developed on rocks by the sand blast is generally somewhat dull, but on certain fine-grained hard rocks such as quartzite it may be highly lustrous.

Transportation by Wind. *Methods of Transportation.*—The method by which rock particles are carried by the wind varies according to size, shape, and density of the particles and with the velocity of the wind. In general, dust particles are carried in suspension; and sand grains by traction, chiefly by rolling. Small, angular grains of dust with relatively large surface areas, as compared with their volumes, remain suspended in air better than larger grains or grains with smooth surfaces which approach the spherical form, and light materials are suspended in air more readily than heavy ones. In ordinary winds sand grains are too heavy to be carried in suspension so that normally they are moved by rolling along the surface of the ground or in part by bouncing, skipping, and gliding. A light zephyr can carry dust in suspension; a gentle breeze can roll fine sand; whereas a strong breeze, with a velocity of about 25 miles per hour, can move sand grains a millimeter in diameter, and gales and hurricanes can carry sand in suspension to heights of hundreds of feet and can roll along the ground pebbles two or three inches in diameter.

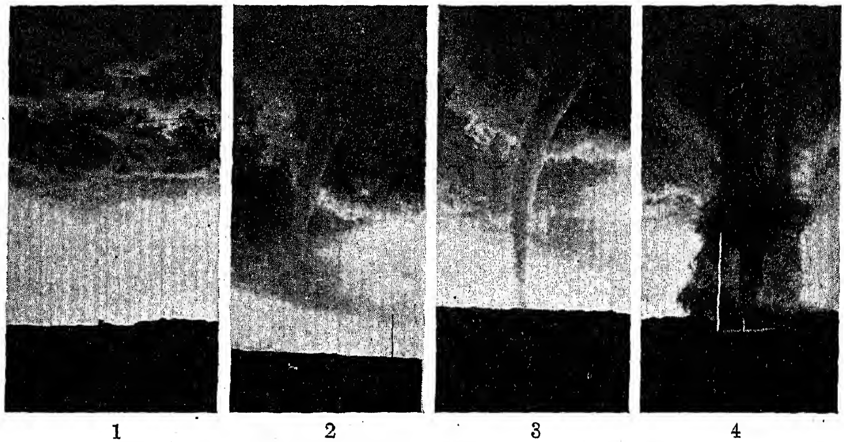

FIG. 53.—Tornado which occurred near Gothenburg, Nebraska, in the autumn of 1930. 1, the tornado cone forming in the clouds; 2, the fully developed cone approaching the earth; 3, the cone as it reached the earth; 4, the cone striking a farmhouse which appeared to explode. (*Courtesy U. S. Weather Bureau.*)

Tornadoes with their great spiral, chimney-like updrafts (Fig. 53) have been known to lift heavy objects and to transport them several miles.

Sources of Load.—The sources of the wind's load of dust and sand are varied. Probably the principal source is the rock waste formed by weathering and corrosion. Disintegrating sandstones, flood-plain and sand-bar deposits of rivers, glacial moraines, beach sands, deposits of dried-up lakes, and the like commonly serve as immediate sources. In addition, volcanic explosions supply tremendous quantities of light, highly angular rock dust which is well suited to the capacity of the wind, but such explosions are infrequent, though temporarily they may be of great importance.

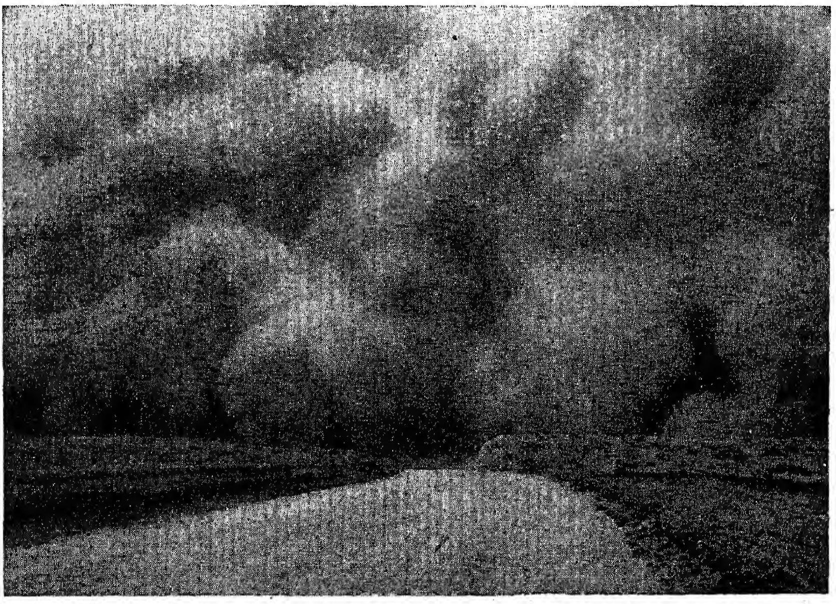

Fig. 54.—An approaching dust storm, western Oklahoma. (*Copyright photograph by Pictures, Inc.*)

Extent of Wind Transportation.—Dust and fine sand (Fig. 54) may be carried to great distances by the wind. Volcanic dust from the explosive eruption of Katmai volcano in Alaska in 1912 was spread to such an extent that the material upon settling formed a deposit a foot thick at a distance of 100 miles to the leeward from the source, and appreciable quantities were carried as far as Seattle, Washington, about 1,600 miles away. When the volcano Krakatoa in 1883 blew more than a cubic mile of its top off and greatly reduced the island on which it stood, dust was carried into the upper atmosphere so as to cause brilliant sunsets for a period of several months, at first near by and about two weeks later entirely around the globe. Certain ancient deposits of volcanic ash, now consolidated into rock, likewise testify to the power of wind transportation. Many examples of dust falls and dust-laden rain- or snowstorms are on record. Thus snow which fell in Minnesota and Wisconsin in the early spring of 1925 carried reddish silt apparently derived from the southwestern states.

Near the sources of the dust much greater quantities are transported. Thus in the drier parts of the prairies and plains which have been termed "the dust bowl" the air is so filled with dust occasionally (Fig. 54) that one can see but a few feet. Driving an automobile is almost impossible. Lights are required in the houses, and even breathing is difficult. The dust piles up at fences, even obliterating them, while around the buildings great dune-like heaps of dust are formed (Fig. 55).

FIG. 55.—A semiarid area in the southwest. Prolonged drouth has converted top soil into dust which has been deposited, partially covering buildings and fences. (*Copyright photograph by Pictures, Inc.*)

This condition has been so discouraging to the inhabitants that during the recent dry period much farm land was abandoned. After one such dust storm, observations on the quantity of dust left over the surface, indicated that 125 tons had fallen on every square mile of the Minneapolis-St. Paul district at a distance of approximately 500 miles from the probable source of the dust.

Dust storms originating in the Sahara desert have been observed to drop silt in Italy on the following day and in France, England, and Germany, 200 to 2,500 miles from the source, a day or two later. Such a dust storm from March 9 to 12, 1901, is estimated to have deposited about 1,960,420 tons of dust in a sheet about 0.25 millimeter thick over Europe and 1,650,000 tons in northern Africa. Likewise ships in the middle and south Atlantic ocean have experienced falls of silt and fine sand, apparently derived from the Sahara desert, and ships off the coast of Japan have received falls of dust from the interior of China, 1,000 miles away. New Zealand has received dust from Australia, 1,400 miles away.

The transportation of sand rolled on the ground is much more limited —generally to distances of a few miles or less. In France, however, sand has been blown inland from the seashore fully 5 miles, and great sand deposits of the Sahara desert lie on a limestone plateau 100 miles from the outcrops of sandstone whose disintegration apparently has supplied the sand.

Sorting and Rounding during Transportation.—The wind's power to transport by suspension is notably different from its capacity for rolling, so that a separation of rock particles occurs depending on the method of transportation. In general, the separation removes clay dust and silt from the larger and heavier sand. The rolling sands themselves may be further sorted on the ground according to the effective velocities of the wind. Pebbles generally are too big to move. Sand particles rolled along the ground are subject to considerable wear so that they become well rounded, even those of very fine sizes (0.3 to 0.1 millimeter or less), whereas suspended dust particles are little changed.

Deposition by Wind. *Dust Deposits.*—Slackening of the wind allows its suspended load to settle slowly out of the air (Fig. 55). Rain and snow are even more effective in rapidly clearing the air. The general haziness of the atmosphere in the Pacific Northwest in late summer is due to dust and smoke from forest fires. It disappears when the fall rains begin. Occurrences of "mud rains" and dust-colored snows already have been referred to. Normally the rate of deposition of dust is very slow, except in connection with volcanic outbursts. At Kodiak, Alaska, in 1912, 5 inches of volcanic dust from Katmai fell during a single night.

Dust deposits in general are widespread and without special form and hence escape much notice, but their total bulk is undoubtedly large. Ancient cities of the Near East were partly buried under such a mantle of dust, apparently on account of a change of climate in the region. Wind-blown deposits in general are called *eolian* deposits. Some deposits of wind-laid dust are of such magnitude as to deserve a special name. To these the term *loess* is applied (Fig. 56). Dust that is composed of volcanic materials is volcanic ash. Prominent loess deposits occur in the Mississippi valley, in the Palouse hills of eastern Washington, in the plains of Germany, in the interior of China, in the pampas of Argentina, and elsewhere.

The typical loess of the Mississippi valley is non-stratified yellowish silt, intermediate in texture between clay and sand, and is composed of a variety of minerals including quartz, feldspar, hornblende, calcite, and many others. The deposits range in thickness from a few feet or less to at least 100 feet. When exposed in the banks of gullies or excavations, the loess reveals the curious property of standing in vertical cliffs even though the material is not cemented together. This is seen at Council Bluffs, Iowa. Apparently the grains are sufficiently angular to interlock rather than roll or slide over each other as most sand grains do. Many exposures of loess show prominent vertical jointing. On account of its texture and the freshness of the component minerals, loess makes fertile, easily worked soil, as it does in the "corn belt" of Iowa and Illinois. The loess of the Mississippi valley lies principally just beyond or on the outer rim of the glacial deposits of the recent ice age, from which deposits the materials of the loess presumably were derived before vegetation had become reestablished. Some loess, however, appears to be related to the valleys of the Missouri and Mississippi rivers. From the periodically dried-up flood plains of these rivers at the present time strong westerly winds whip up the dust and carry it to the adjacent uplands.

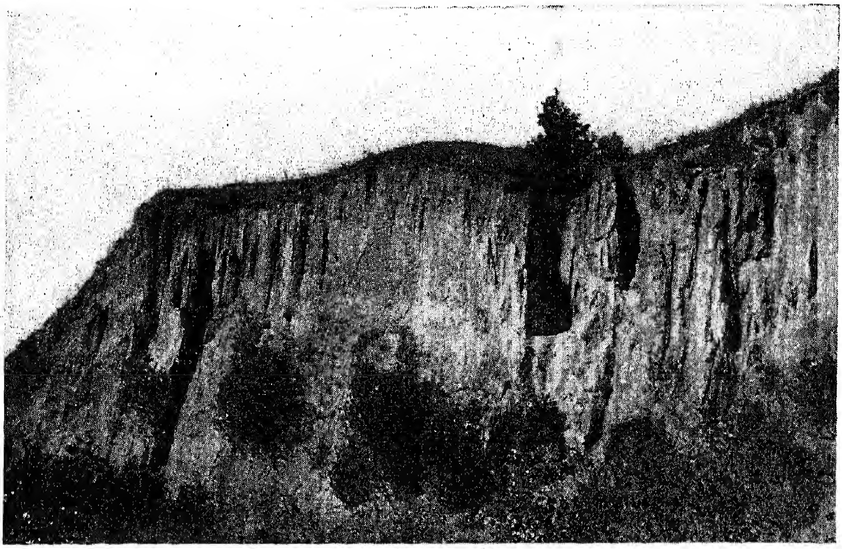

FIG. 56.—Fifty feet of loess showing a nearly vertical face at Missouri Valley, Iowa. (*Photograph by Alden, U. S. Geol. Survey.*)

The loess deposits of China are said to reach the enormous thickness of 1,000 feet in Shensi and adjoining provinces. This loess is believed to have accumulated from silt blown in from the deserts of Central Asia by the prevailing westerly winds. It is so easily eroded by wind and rainwash that certain roadways in it, through the wear of centuries of travel, have become depressed into deep, narrow, canyon-like defiles. To restrain soil erosion, the loess-covered valley slopes are terraced by the Chinese farmers. Many of the farmers in the loess district inhabit caves excavated in bluffs of the loess.

Volcanic ash is finely broken volcanic material, and much of it contains sharp, angular grains of glass commonly inclosing steam bubbles or showing curved edges where the bubbles were broken. Volcanic tuff, composed of volcanic fragments partly consolidated, is a common rock in certain sections of the western part of the United States and has been found among older rocks at many places on the earth.

Dunes.—Eolian deposits of sand may be of very irregular shapes, but definite hills or ridges of sand heaped up by the wind are more common. These wind-built sand hills and ridges are dunes (Fig. 57). They are formed in much the same manner as snowdrifts. They are started by some obstruction such as a bush, boulder, fence, or other obstacle which causes an eddy or otherwise thwarts the sand-laden wind. Once begun, the dunes themselves offer further resistance so that they tend to grow. The addition of sand is partly on the front slope but especially beyond the crest on the leeward side. The source of the sand is usually (1) a sandy beach, as on the shore of Lake Michigan at Dune Park, Indiana, or along the coast of France on the Bay of Biscay; (2) a sandy river plain, as along the lower Columbia River or along the Arkansas River in western Kansas; or (3) a disintegrating sandstone, as in western Nebraska and in the Sahara desert.

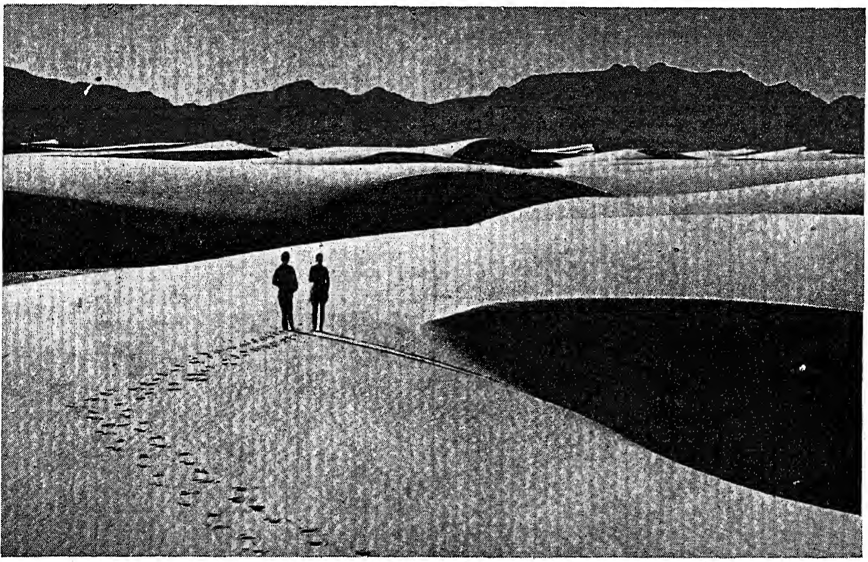

FIG. 57.—Dunes of gypsum sand in White Sands National Monument, New Mexico. The San Andreas Mountains are shown in the background. (*Photograph by Geo. A. Grant, courtesy U. S. Department of the Interior.*)

Dunes take various shapes (Fig. 58), according to the source of the sand, the amount available, and the velocity and constancy of direction of the winds. They may be roundish hills, elongate ridges, crescents, or hummocks quite irregular in outline. In profile, dunes normally show on the windward side a long, gentle slope, usually 5 to 15 degrees from the horizontal, and on the leeward side a steeper slope, usually 15 to 25 degrees and rarely more than 30 degrees. The highest possible angle is the angle of rest of loose, dry sand. Large dunes may be 300 feet high with front and back slopes of 10 and 25 degrees respectively and nearly half a mile long. Coastal dunes, well fed from beaches and moving inland under fairly regular wind conditions, commonly take the form of a succession of nearly parallel ridges generally 10 to 50 but at places 200 to 300 feet high. Inland, or true desert, dunes formed in relatively open country from a limited or moderate supply of sand commonly take horseshoe or crescent forms, known as *barchanes;* the crescentic shape is due to sand blowing around the sides as well as over the top of the pile. The horns of the crescent point in the direction of the movement. Many African dunes of this type reach heights of 200 to 300 feet, and some of the largest are said to be 1,000 feet high. If the winds are very strong and the supply of sand not too great, the dunes may be drawn out into narrow strips or ridges trending in the direction of the wind. If the supply of sand is large, barchanes may coalesce into transverse ridges with irregular crests. Variations of directions of wind and resistance offered by bunches or clumps of vegetation may make dunes very irregular in shape.

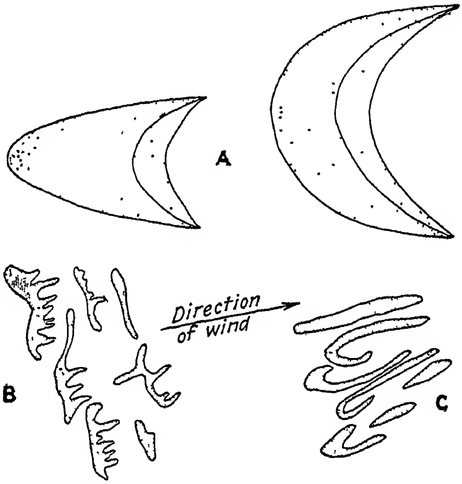

FIG. 58.—Diagrams showing the relationships in form and in orientation of dunes to the supply of sand and to the strength of the wind. *A*, barchanes formed by small supplies of sand and moderate winds; *B*, transverse ridges formed when supply of sand is large and winds moderate; *C*, dune ridges formed with large sand supply and violent winds. (*After Cornish.*)

A dune moves by the transfer of sand from the windward to the leeward side (Fig. 59). The wind rolls sand up the gentle slope to the crest of the dune, where in the eddies the sand drops out of reach of the wind and rolls to a position of rest. More and more sand, blown from the face of the dune, is carried over so that eventually the positions of the crest and the extremities of the dune are shifted to leeward (Fig. 60). On account of the large amount of sand to be transferred, the rate of dune movement generally is slow, perhaps only a few inches or a few feet per year. Dunes in Germany have been known to pass entirely over certain points in about 60 years (Fig. 61). Exceptionally, dunes move hundreds of feet per year; some small ones in Asia are said to be blown across country as much as 60 feet in a single day.

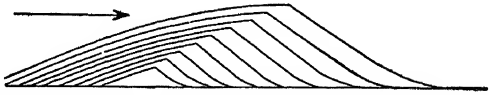

Fig. 59.—Cross section of a dune. The crest has advanced in the direction of the wind, but the dune as a whole has not migrated.

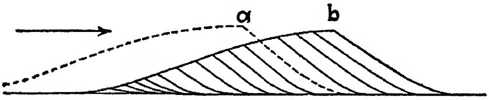

Fig. 60.—Cross section of a dune that has migrated from *a* to *b* by the shifting of the sand from the windward to the leeward side of the dune.

Even a slow migration of dunes is sufficient ultimately to overwhelm forests (Fig. 62), buildings, or other objects which stand in the way, so that locally the control of dunes is of economic importance. To stop their movement the planting of certain hardy grasses, shrubs, and trees, adapted to sandy soils, on the windward slope of the dunes has been found effective, especially on the French coast. Where the expense is

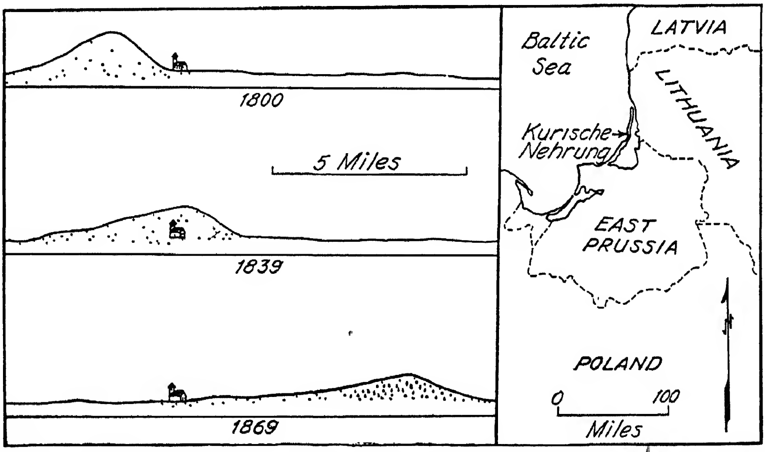

Fig. 61.—Diagram showing how the town of Kunzen on the Kurische Nehrung, East Prussia, was buried by a sand dune after the year 1801 and subsequently in 1869 was exhumed by the migration of the dune. (*Based on diagrams by Berendt.*)

justified, a coating of heavy oil may be used to hold the sand in place. Commonly it is easier for man to avoid the dunes than to stop them. Some dunes have become fixed naturally as a result of changes in climate, cessation of sand supply, or growth of vegetation. Holes in fixed dunes, scoured out by the wind where the vegetal cover has been rent, are termed "blow-outs."

Fig. 62.—Drifting sand covering a forest.

The surfaces of dunes commonly are rippled into little ridges and furrows, called ripple marks, which are like dunes in miniature. They are the result of friction eddies, which may be likened to roller bearings between the wind and sand.

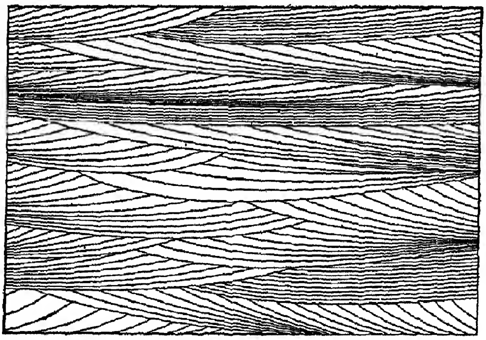

Fig. 63.—Eolian cross-bedding in sandstones.

Winds of different velocities at different times blow sand grains of differing sizes over the dunes, so that the sands deposited on the back slope show sorting into thin layers of coarse and fine sand. Variations in the amount and direction of slope of the leeward surface and shifting of directions of the wind and hence of the sites of deposition cause these beds to slant at different angles and in various directions within short distances, so that amid the complexities of scour and deposition accompanying dune movement a maze of inclined bedding or cross-bedding is produced in the interior of dunes. The great extent and irregularity of this cross-bedding in dune sands are so characteristic that the sand of certain ancient sandstones which show such structure as in Fig. 63 is believed once to have been in dunes.

Most dune sands consist largely of quartz with small amounts of other minerals, but notable variations occur on account of differences of weathering and sorting in the previous history of the sand grains. The beach dunes of Bermuda, composed mostly of calcite, and certain dunes in Otero County, New Mexico, made of nearly pure, snow-white gypsum, are exceptional (see Fig. 57).

It is noteworthy that much eolian sand is not in the form of dunes but is variously disposed as sheets or as drifts against mountain slopes, in the lee of cliffs, etc. Furthermore, the actual dune areas in deserts are relatively not large. Thus in Arabia, one of the bleakest deserts of the world, not more than one-third is sand, and of the great Sahara desert not more than one-ninth is occupied by dunes. In the desert areas of the United States the ratio is very much less than one-ninth. Expanses of relatively barren rock are more characteristic of deserts than are dunes.

The sand of dunes is generally well sorted and contains little coarse material. At places where lightning strikes loose sand, the partially fused material forms small rods and tubes of glassy material which may easily be picked out of the sand. These rods and tubes are fulgurites (Fig. 64), and they may form in any loose sand.

FIG. 64.—Fulgurites formed by melting of sand where lightning struck the ground.

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CHAPTER IV

GROUND WATER

The waters of the earth include those of oceans, lakes, and streams, the vapor in the atmosphere, and the ground water that is contained in the openings in the earth's crust. Of the water that is precipitated upon the earth a part is carried by the streams and rivers to the sea, another part is evaporated, and still another part soaks into the ground and is stored in the openings in rocks. This latter portion is *ground water* (Fig. 65).

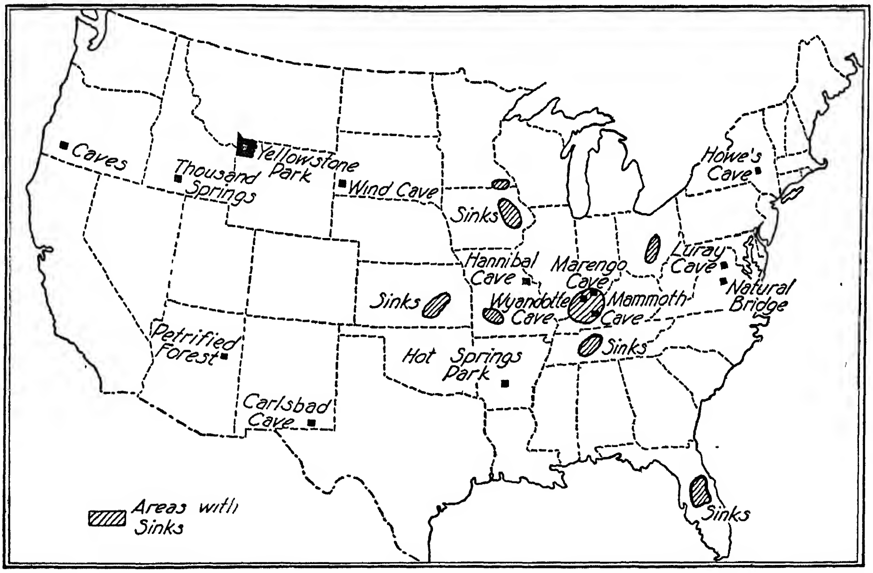

Fig. 65.—Map of the United States showing the locations of features of special interest produced by the action of ground water.

Meteoric water is that which falls upon the surface of the earth. Probably part of the meteoric water was once a part of the magmas or molten rock matter that at various times has been thrust from below into the crust of the earth. It is probable that waters of certain hot springs are in part of magmatic origin. The waters of these join the run-off, are evaporated, and later are precipitated. Thus magmatic, or juvenile, water becomes meteoric water. Connate water is water that was trapped in the sedimentary beds when they were deposited. The salt water that is found along with oil in the productive beds of many petroliferous areas is believed to be connate water. Such water probably has remained in the strata in which it is found since the time when the beds that contain it were laid down in the sea.

The total annual fall of meteoric water upon the earth's land surface is estimated to be about 26,679 cubic miles. Its distribution is irregular. In the polar regions the annual snowfall represents from 8 to 15 inches of water, whereas on certain southern slopes of the Himalayas the rainfall may be as much as 500 inches per year. In other regions such as the vast desert areas of northern Africa, Central Asia, Australia, and along the coast of Peru very little water is precipitated. Among the Canary Islands are localities that have practically no rainfall for long periods.

Descent of Ground Water.—The amount of rain water that penetrates the earth is determined by several factors, the chief of which are stated below.

1. Amount and kind of precipitation. Within certain limits the amount of water that soaks into the ground is determined by the amount of precipitation as rain. In desert areas ground water generally lies deep, and little water occurs near the surface, because there is little rain.

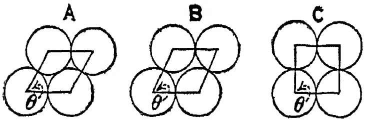

FIG. 66.—Diagrams illustrating pore space resulting from different arrangements of sand grains. *A*, showing the least pore space; *B*, more pore space; *C*, most pore space. (*After Meinzer, U. S. Geol. Survey.*)

2. Rate of precipitation. The more rapid the fall the less water sinks into the ground, for the surface soon becomes saturated. The same is true of the melting of snow—the more rapid the rate of melting the less water sinks into the ground, especially into frozen ground.

3. Slope of the surface. The steeper the slope of the ground the greater the percentage of the run-off. The flatter the ground is the more water will sink below the surface, because the run-off is retarded and the water has a longer time to soak into the ground.

4. Porosity of the soil and rock. Weathered and stratified rocks are usually more favorable for the entrance of water than massive, igneous rock. The part of a rock which is occupied by voids determines its porosity (Figs. 66, 67, 68). Thus if a gallon of sand will hold 0.3 gallon of water when saturated, the porosity is said to be 30 per cent, for three-tenths of its volume is made up of pores between the grains. The porosity of different types of rock varies from less than 1 per cent in massive granite to more than 40 per cent in poorly cemented sandstones. The granite selected for the sarcophagus of the tomb of General U. S. Grant was considered the strongest granite in the United States. Poros-

ity tests showed that it possessed about one-fourth of 1 per cent pore space. Thus even the strongest and most massive rocks contain measurable pores.

5. Structure of the rock formations. Inclined strata will allow more water to penetrate the earth than flat-lying beds. Water passing down inclined beds will follow the most porous layers. If the rocks are horizontal, the water passing downward must cross also the least porous beds.

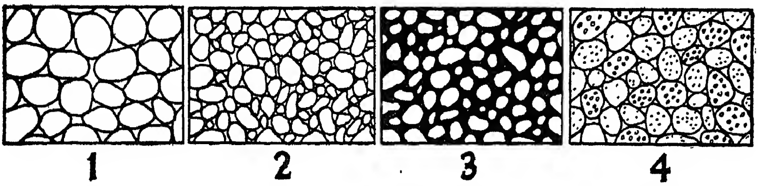

Fig. 67.—Diagrams showing various factors that influence the porosity of sedimentary rocks. 1, a well-sorted sand having high porosity; 2, a poorly sorted sand having lower porosity; 3, a poorly sorted sediment with its porosity reduced by the deposition of mineral matter in the interstices; 4, a well-sorted sediment consisting of pebbles that are themselves porous, thus increasing the porosity of the rock as a whole. (*After Meinzer, U. S. Geol. Survey.*)

6. Amount and kind of vegetation on the surface. Plants and organic matter derived from plants check the flow of surface water, and more water sinks below the surface. Forests and meadows hold back the run-off and also retard evaporation.

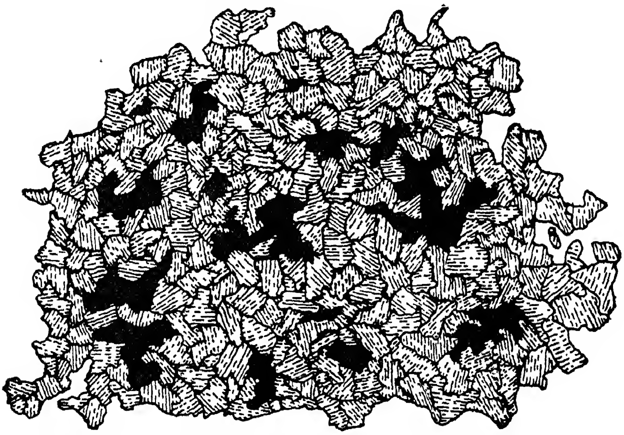

Fig. 68.—Diagram showing openings (black) in limestone formed by solution.

7. Amount of moisture in the atmosphere. If the humidity is low immediately after a shower, more of the rainfall evaporates before it can sink into the earth. This is especially true in arid regions, where even after heavy rains the bulk of the water dries up or by evaporation passes again into the atmosphere.

Level of Ground Water.—The terms "ground-water level," "water table," and "plane of saturation" are commonly used to describe the upper surface of the zone within the earth below which the openings in rocks are filled with water (Fig. 69). The upper limit of this zone of saturation is not a plane but is undulating. It tends to follow the undulations of the topography of a region, but it is more regular. It generally lies lower below a valley than below a hilltop, but it lies deeper below a hilltop than below a valley.

The level of the plane of saturation is controlled by several factors, such as the amount of rainfall, the amount of evaporation, and the porosity of the rocks. The general level does not respond at once following each period of rainfall because of the time required for the water to percolate downward through the unsaturated materials above the water table. In fact during a heavy rainfall a saturated zone may be formed near the surface, constituting a temporary "perched" water table at some distance above the plane of general saturation.

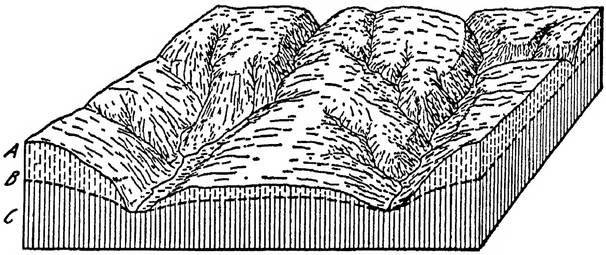

Fig. 69.—Diagram showing the position of the ground-water table in an area of undulating topography. *A*, zone of leaching or weathering; *B*, ground-water table; *C*, zone of saturation.

The rise of the water table following a period of rainfall never corresponds in amount to the number of inches of precipitation, owing to run-off, losses through evaporation, absorption by vegetation, and absorption by decomposed rock materials in the unsaturated zone above. The difference of elevation between the top of this zone in a wet year and in a dry year is normally greater under the hilltops than on the slopes and in the valleys. Since the water table oscillates with climatic changes, there is a zone within the earth that is above ground-water level in dry periods but below it during wet periods. In moist, hilly regions this zone may be of considerable vertical extent.

Movements of Ground Water.—Above the water table, ground water moves downward and generally not far laterally. In the zone of saturation it is not stationary. Its movements are slow, but it tends to migrate along the paths of least resistance through the rocks; and if there is a lower outlet along the bottom of a valley, lake, or other basin, it will move to it even though it may follow a very crooked route before it finds a point where it can issue again at the surface. If the paths of least resistance are downward, the water may sink to great depths before it rises through some porous formation or fracture which crops out at a point lower than that at which the water first entered the saturated zone. The cause of movement is gravity.

The downward movement of water toward the zone of saturation is termed the "vadose," or shallow circulation. The thickness of this zone is variable, since its lower limit is determined by the position of the ground-water level. Near permanent streams or lakes or other bodies of water it lies near the surface. In hilly regions with average rainfall, however, its depth from the surface varies from a few to several hundred feet. In arid regions, where the amount of rainfall is small and evaporation rapid, this zone may extend to much greater depths.

Below the water table the circulation of water depends on the relief of the region and on the number, continuity, spacing, and size of the openings in the rocks. As a rule, the flow of water in this deeper zone is much slower than in the vadose zone, because the openings are less numerous and smaller. For this reason friction on their walls is greater. In fine-grained rocks the underground circulation becomes exceedingly sluggish, the water moving perhaps not more than a few feet per year. The depth to which the surface waters penetrate varies with the character of the rocks. In some rocks, surface waters reach a depth of several thousand feet, whereas in others very little water is collected at depths of more than a few hundred feet. In the copper-bearing rocks of Keweenawan Point, Michigan, shafts have been sunk over a mile below the surface, and at a number of places the rocks are dry and dusty near the lower ends of the shafts. Many holes bored deep into the earth in search of oil have penetrated dry rocks at depths of 2,000 or 3,000 feet. A study of the movement of underground water indicates that there is an indefinite division between a sluggish deep circulation and a zone of essentially stagnant waters at still greater depths. Where these deeper waters are charged with salts similar in composition to those of sea water, they are considered residual sea waters or connate waters which have remained in the strata since they were deposited.

Springs.—Springs are formed wherever underground waters flow to the surface through natural openings in the ground. The rate and manner of flow are regulated by the geological structure of the mantle rock and of underlying formations (Fig. 70). Ground water always flows along planes or channels of least resistance. At first it percolates and seeps slowly through the rocks, but in time it wears well-defined courses. Springs usually issue upon a hillside or in a valley. An ordinary hillside spring is formed where sand, gravel, sandstone, or other porous strata rest upon impervious beds. Where the water comes to the surface along an escarpment it "weeps" out in the form of hundreds of small seepage springs. If the strata are inclined, deep-seated fissure springs or artesian springs may issue through points along fault planes that cut the impervious strata. Such fissure springs may discharge fresh water on the floor of the sea, where it will rise through the heavier salt water before the two become mixed. Such springs are found along the coasts of the Mediterranean Sea. In the Gulf of Argos, Greece, a body of fresh water estimated to extend over an area 50 feet in diameter, probably represents the exit of a fissure spring, for it discharges fresh water with such force that it forms a convex surface on the sea.

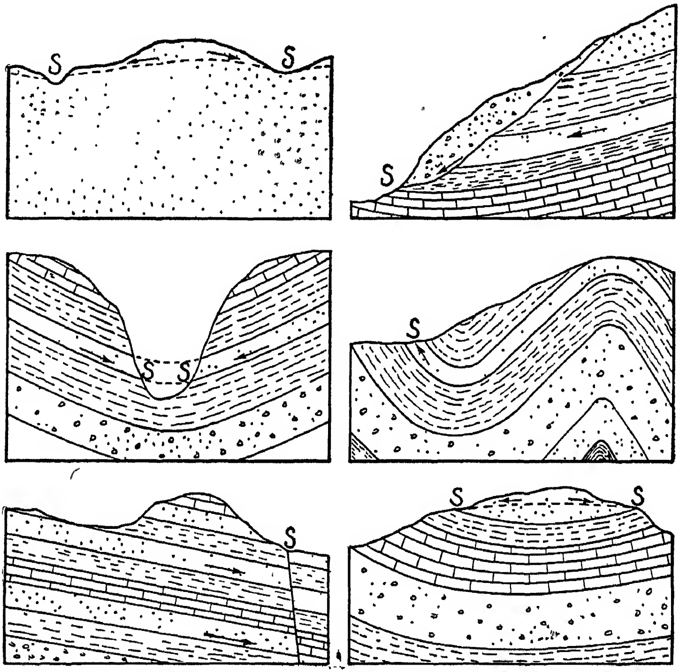

Fig. 70.—Diagrams showing various kinds of subsurface rock structures favorable for the formation of springs. Locations of springs are marked by the letter *S*.

Hot Springs.—Hot springs are vents in the earth's crust from which hot water issues. Some hot springs and a considerable number of warm ones are found in areas remote from igneous centers, and it is believed that the water of such springs is normal rain water that has penetrated the ground and moved downward to great depths. The temperature of the earth's crust increases downward at a rate of about 1°C. for every 30 meters (about 100 feet). Ground waters at a depth of a mile would be about 50°C. hotter than the average surface temperature; and if such water were to rise without much dilution, it would be noticeably warmer than surface waters. Many hot springs and the hottest ones are found in volcanic areas. This suggests a connection between certain hot springs and cooling igneous rocks. Gases like those that issue from the surface in volcanic areas are present also in many hot springs, and certain vents from which gases issue in dry seasons become hot springs in wet seasons. In some the gases issue vigorously and agitate the water. Such are known as boiling springs. Some that contain rock fragments, particularly the oxidized particles of iron, color the water yellow or red and are called "paint pots," "ink bowls," etc. Algae, which are simple forms of vegetable life, thrive in the warm waters of certain springs. Some of them are brightly colored, and they color the walls of the springs.

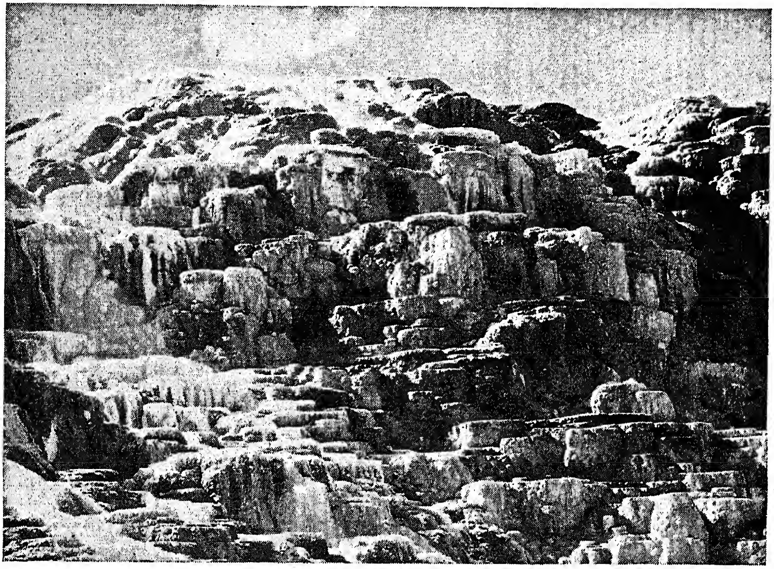

FIG. 71.—Jupiter Terrace, Mammoth Hot Springs, Yellowstone National Park. The deposits are formed by mineral-bearing hot waters that overflow the basins and build up the sides of the pulpit-like terraces. (*Courtesy Union Pacific R. R.*)

FIG. 72.—Crater of Oblong Geyser, Yellowstone National Park. (*Photograph by Haynes.*)

Waters and Deposits of Hot Springs.—The waters that issue from hot springs contain many salts, chiefly alkalies and alkaline earths, which are present as carbonates, chlorides, and sulphates. Certain warm springs in areas remote from volcanic activity carry abundant mineral salts, but alkalies and chlorides are much less prominent in such waters than in those of volcanic areas. Tufa is material deposited by springs around their vents. It is generally deposited more abundantly by hot springs, because their waters before issuing can carry more mineral matter in solution than colder water. Much tufa consists chiefly of calcium carbonate. Deposits so formed of nearly pure calcium carbonate are called "travertine." If the calcium carbonate forms crystalline bands and takes a good polish, it is "Mexican onyx." When the tufa contains fragments cemented by calcium carbonate, it is a "breccia," and some of it is highly prized as an ornamental stone. Spring deposits also are called "sinter" and, if of calcium carbonate, "calcareous sinter." *Geyserite* is a term applied to siliceous hot-spring deposits, particularly

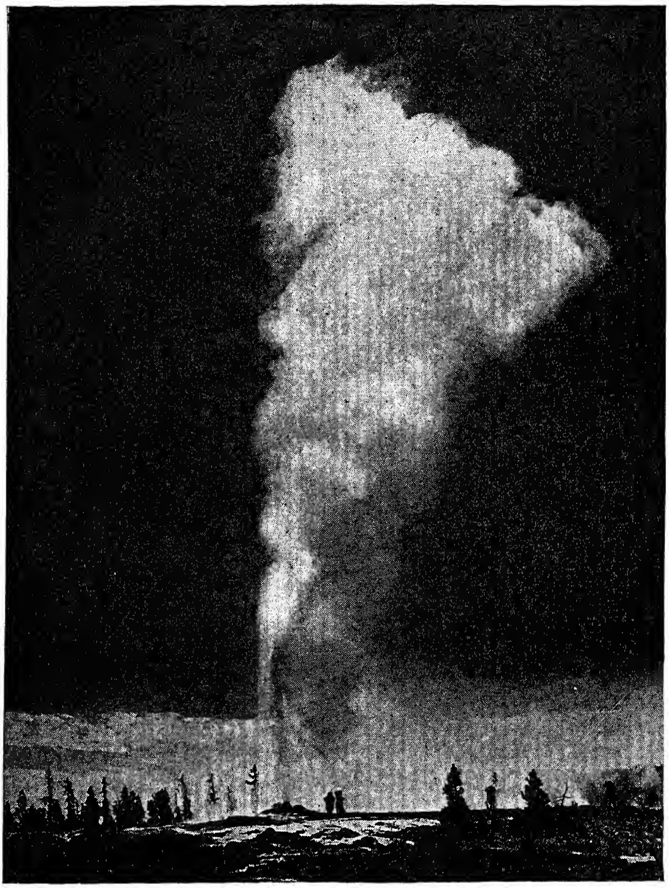

FIG. 73.—Old Faithful geyser in eruption, Yellowstone National Park. (*Courtesy Union Pacific R. R.*)

to deposits made by geysers. Where deposits are abundant, a spring or geyser may build a mound or terrace (Fig. 71) or a bowl-like structure around the orifice of the spring or geyser, and the water that issues may pour over the top of a bowl (Fig. 72). Deposition of long-continued overflow builds up the terrace. Calcium carbonate is the salt most abundantly deposited by both hot and cold springs. It frequently forms great beds near the outlets, and at resorts it clogs the pipes that lead the hot water to bathing pools. Silica also occurs abundantly in many hot waters, and it is the chief material deposited by some of the famous hot springs of Yellowstone National Park. If iron oxide is deposited by the water along with other materials, the spring is a chalybeate spring.

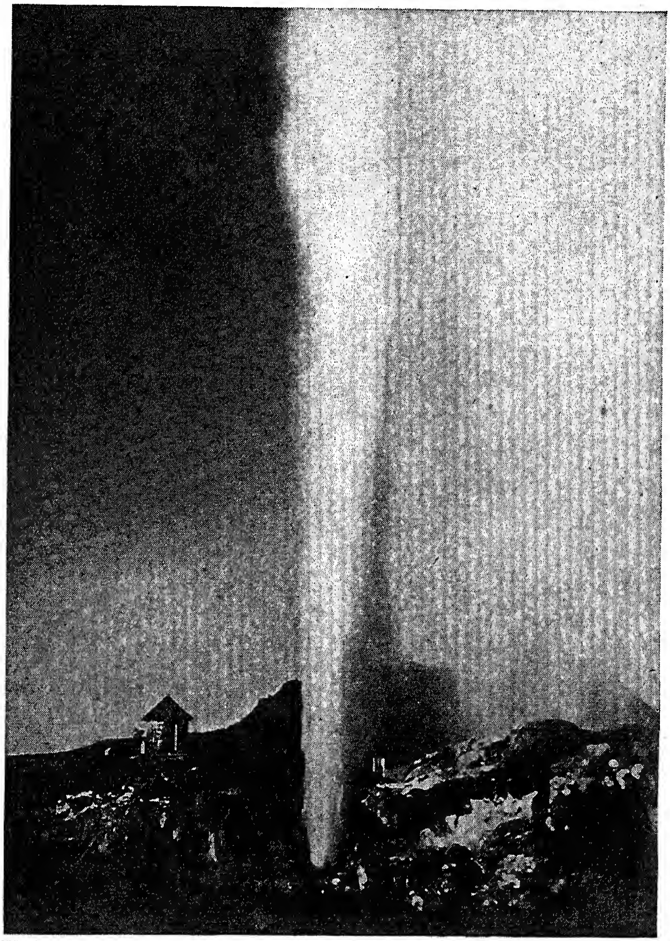

FIG. 74.—Wairoa geyser in eruption, New Zealand. (*Photograph courtesy New Zealand Government Publicity.*)

Geysers.—Geysers are hot springs from which the water is expelled vigorously at intervals. They are much less numerous than ordinary hot springs. Groups of geysers are situated in Yellowstone Park in the United States, in Iceland, and in New Zealand. These areas are regions of recent volcanic activity. Certain geysers, when they erupt, throw hot water several hundred feet into the air. In others the water reaches only a few feet above the surface. Most geysers are active at irregular intervals. In some the eruptions occur many days apart; in others the intervals are weeks or months. In general, the water of the geysers does not differ from that of other hot springs. In certain geysers, after the activity ceases, some of the water flows away from the vent, but in others practically all of it flows back into it again. One of the best known geysers is Old Faithful of Yellowstone Park (Fig. 73), which for many years erupted regularly at intervals of less than 60 minutes. In recent years this interval has changed considerably, yet the geyser is still comparatively regular in its eruptions. An eruption of Wairoa geyser, New Zealand, is shown by Fig. 74.

The eruption of a geyser is preceded by rumblings and by violent boilings. The water flows over the top of the vent, and soon low columns of water appear. These are followed by strong jets, which are thrown scores of feet into the air. A record of an eruption of Old Faithful is shown by Fig. 75. The numbers on the horizontal line at the base of the figure show the minutes after eruption began, and the vertical lines show the heights to which the water jets rose. These were obtained by sighting and reading the angle of sight to the top of the water column. Around the base of the column there was much steam, and it was not possible to record the height of the erupted column when it did not rise above the steam. The eruption lasted about 20 minutes and increased in violence. At the end of 15 minutes the water began to rise higher, and it reached its greatest height about 16 minutes after the activity began. Subsequently the eruption subsided, the water was thrown to lower heights, and finally it died down and stopped.

Causes of Eruptions of Geysers.—The water thrown from geysers is mainly rain water that has soaked into the ground, and the heat is supplied by hot lavas or other igneous rocks. The gases, other than steam, and some of the waters probably have originated in the cooling igneous rocks. The theory of eruption generally accepted is that of Bunsen, who studied the geysers of Iceland. The temperature at which water boils increases with pressure. This temperature is 100°C. at the surface of the earth, where the pressure above the water is 1 atmosphere, or about 14.7 pounds per square inch. At 2 atmospheres it boils at 120°C., and at 10 atmospheres at 180°C. A column of water 33 feet high has a pressure of about 1 atmosphere; that is, a column 1 square inch in cross section and 33 feet long weighs about 14.7 pounds. The temperatures

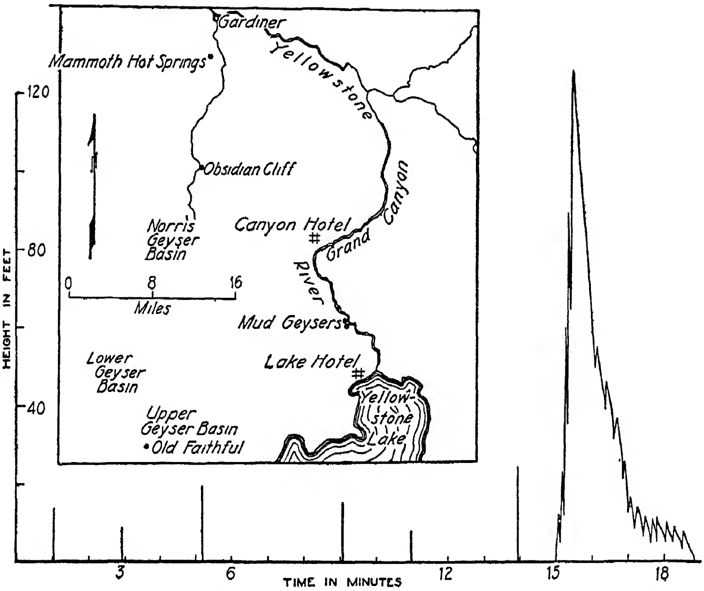

FIG. 75.—Diagram showing an eruption of Old Faithful geyser. The height to which the hot water is thrown upward is shown by the line at the left of the figure. The times of the eruptions are shown on the horizontal line at the base of the figure. The measurements were made by sighting at the top of the water column. There was so much steam during the early eruptions that only maxima of the jets could be seen. After the eruption began, it was more than 15 minutes before the highest eruption was attained. The inset shows location of Old Faithful geyser. (*After H. E. La Tendresse.*)

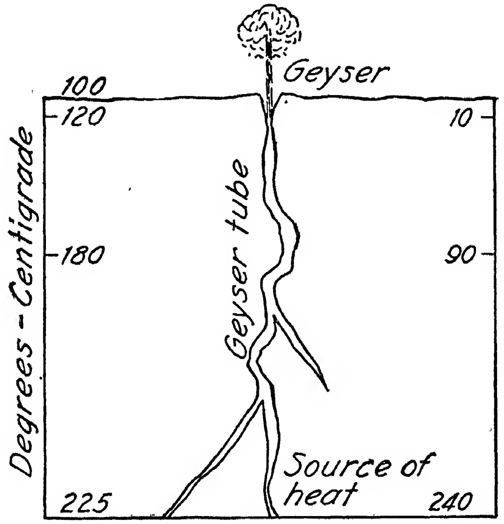

FIG. 76.—Diagrammatic cross section illustrating geyser eruption according to the theory of Bunsen. The numbers at the left of the diagram show boiling points of water at certain pressures corresponding to the weight of a column of water at certain depths, in meters, shown at the right side of the figure. (One meter equals 3.28 feet.)

at which the water will boil at various pressures are shown in the table below and in Fig. 76.

If water flows into a fissure or tube (Fig. 76) and becomes warm with depth by absorbing hot gases or by contact with hot rocks, it may remain as water, although it is much hotter than 100°C., which is the boiling point at the surface. It does not become steam, because of the water pressure above it. If water continues to enter the tube, ultimately it will flow over at the surface, and that from the greater depths will rise to take the place of the overflow; but the water at great depth is so hot that it would become steam if it were not for the great pressure of the water column above it. The crookedness of the tube prevents convection. Hence the superheated water is confined to the lower part of the column. Finally the temperature rises to such a point that steam is formed, even under the great pressure. This pushes up the column of water and, mixing with the water, makes it lighter; some of the water spills over at the top and reduces the pressure. Part of the superheated water then flashes into steam, and an eruption follows. It is probable that the superheating of water causes the eruption of some geysers and that the collection of steam and other gases in the high parts of crooked tubes of geysers operates to cause the eruptions of others.

The addition of soap will cause a geyser to shorten the periods between its eruptions. That is probably due in part to the formation of soap bubbles, which lighten the column of water and hasten the time when it is overbalanced by the gas pressure.

APPROXIMATE PRESSURE NECESSARY TO HOLD WATER IN THE LIQUID STATE AT CERTAIN DEPTHS

(Mainly from a Table by Van Hise)

1 Temperature, degrees centigrade	2 Pressure, atmospheres	3 Depth necessary to produce pressure of column 2, meters
100	1	0
120	2	10
180	10	90
225	25	240

Column 1 shows various temperatures of water.

Column 2 shows approximately the pressures at which water boils at the temperatures of column 1.

Column 3 shows depths in a water column where these pressures exist.

Mineral Matter in Spring Water.—Even the clearest and most sparkling spring water contains some dissolved mineral matter. As a rule it consists principally of carbonates of calcium, magnesium, and sodium and sulphates of calcium and sodium, with smaller amounts of silicates, phosphates, and chlorides. Some of these materials are taken into solution through the action of gases absorbed by rain from the atmosphere. Other materials are derived from decomposing rock and organic matter in the soil. The gases are mainly carbon dioxide, hydrogen sulphide, and marsh gas. In areas covered with a mantle of humus, organic acids also are abstracted from the soil and aid in decomposing minerals and in forming soluble salts.

A spring in which the amount of mineral matter in solution is relatively great is a mineral spring. One in which lime predominates is a calcareous spring; and where the water carries a large proportion of iron, the spring is a ferruginous, or chalybeate, spring. Iron-bearing springs are readily recognized by the yellow- or brown-ochre deposits around or near their vents. Springs high in sodium chloride are brine springs. They occur where beds of rock salt exist beneath the surface, where the rocks have chloride minerals disseminated through them, or where salt water issues from some deep-seated reservoir. The term medicinal spring is applied to mineral springs which are supposed to have curative effects in certain types of diseases. Such springs may contain sulphurous waters, alkaline waters, or bitter waters, each of which derives its name from the soluble salt that predominates.

The amount of dissolved mineral matter brought to the surface by spring waters is enormous. Thousands of tons of gypsum issue annually at the springs of Leuk, in Switzerland, and it has been estimated that the famous springs of Bath, in England, bring up so much mineral matter in solution yearly that if it were taken out of the water and made into a monument it would make a column 9 feet in diameter and 140 feet high. In central Florida, Silver Springs carry about 600 tons of mineral matter daily, and Falls Creek, in Oklahoma, receives water from springs that carry so much lime that a series of travertine dams have been deposited across the stream valley.

Wells.—Prehistoric man used the water from springs, brooks, rivers, and lakes. His villages were built where water was readily obtained at the surface. However, with the growth of civilization, large quantities of water were required in regions where surface water was not available or where it was too polluted for domestic use. Digging or boring for water dates back to very early historic times, especially in China and India. In Babylonia irrigation works were constructed as early as 2000 B.C. In India today more land is irrigated from wells than from streams.

Most wells are holes dug or bored into the earth to a point below the water table. They serve as reservoirs into which the ground water percolates. If the water level is near the surface, the wells are shallow;

whereas on high plateaus where the water level is as much as several hundred feet below the surface, correspondingly deep wells are required in order to obtain water. Most wells are sunk until they penetrate a permeable rock below the water table. This often necessitates boring through hundreds of feet of impervious clays or shales that are saturated with water, but the pore spaces in such rocks are so minute that water cannot flow through them rapidly enough to supply the well adequately.

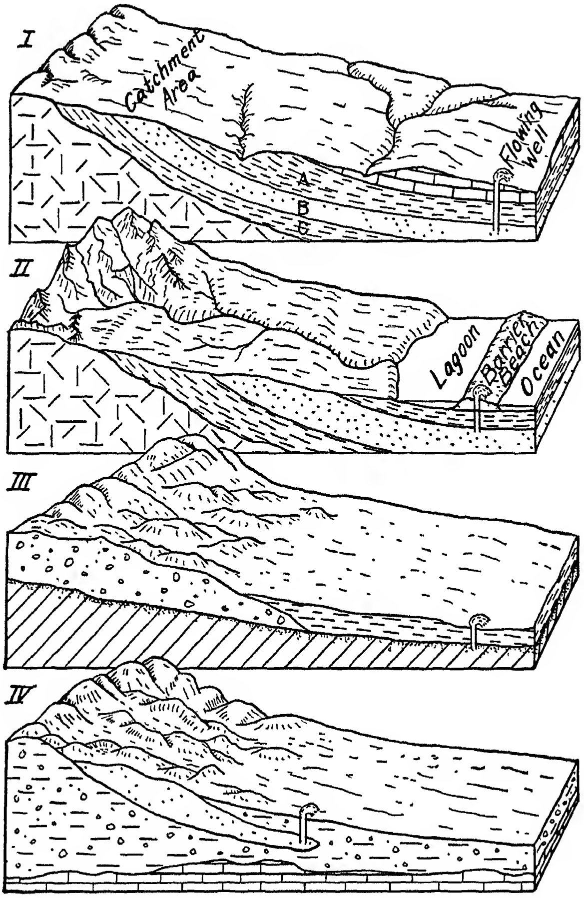

Fig. 77.—Diagrams showing conditions favorable for flowing artesian wells. The porous beds (dotted as stratum *B*, diagram *I*) receive water from the rain in the regions where they crop out (catchment areas). Impervious beds (*A* and *C*) above and below the water-bearing rock. *I*, land area with inclined sedimentary strata; *II*, coastal area where fresh artesian water is obtained on a barrier island which is surrounded by salty sea water; *III*, an inclined fractured and porous zone serving as the reservoir rock; *IV*, lens of sand or gravel in clayey glacial till.

Artesian Wells.—Artesian wells received their name from Artois, a province in France, where the water in many wells rises above the surface of the earth, as it does in fountains. Today the term artesian well is applied to any deep well from which ground water is obtained, even though the water does not rise to the surface. Artesian flow takes place because of differences in the pressure under which ground water exists in different parts of a water-bearing stratum. The principle involved is that expressed in the maxim, "Water seeks its own level." The force causing it to seek that level is gravity. The pressure produced by gravity acting on water is hydrostatic pressure. The water-bearing bed may be compared to a tube filled with water where the intake is higher than the outlet. The conditions essential for a flowing artesian well (see Fig. 77) are summarized below:

1. A pervious stratum which water enters and passes through.
2. An impervious bed above the permeable one to prevent the water from escaping to the surface as springs.
3. An impervious stratum below or else a tightening of the rocks with depth to prevent the water from escaping downward.
4. An inclination from the horizontal of the permeable bed so that the place at which the water enters it will stand at a higher elevation than the surface of the earth at the well. The force of gravity will then cause the water to flow downward. This condition is commonly called "head."
5. The porous stratum must crop out so that water may enter it, or if covered it must be covered by permeable material at the intake.
6. There must be adequate rainfall to supply water.

These required conditions apply only to sedimentary strata. Artesian flows may be obtained from other kinds of rocks even from unconsolidated sediments, but the structural and textural relations must be such that they contain water under hydrostatic pressure. Artesian flows may be obtained from bedding, cleavage, or shearing planes, from solution passages, from joint and fault fractures, and from contacts of sedimentary with igneous or metamorphic rocks.

At many places the sources of the water in the intake areas are scores or even hundreds of miles distant from the wells. In parts of Arkansas, Alabama, California, and Arizona the principal water supplies for irrigation projects are drawn from artesian wells that derive their water from the foothills of the mountainous areas scores of miles away.

Depths of Artesian Wells.—The depth to the porous stratum is determined by the geological structural relations. The Grenelle Artesian in the Paris basin is 2,000 feet deep; a well near Leipzig, Germany, is 5,735 feet; one near Pittsburgh, Pennsylvania, is 4,625 feet deep. Most of the wells in the famous Dakota sandstone basin are 1,000 feet deep or more. Many flowing wells in the glacial drift are less than 100 feet deep.

Chemical Activity of Ground Water.—The various processes of rock decomposition, such as oxidation, carbonation, and hydration, which have already been described as phases of weathering, are active in the zone of circulating ground waters. Not only do they produce reactions within the earth that are analogous to those at the surface, but at many places the waters are more active because of the greater heat and pressure that exist below the earth's surface. The various phases of the chemical work of ground water may be divided into two main groups, namely *solution* and *deposition*. These two processes may be going on simultaneously, for ground water may be dissolving one mineral at the same time that it is depositing another.

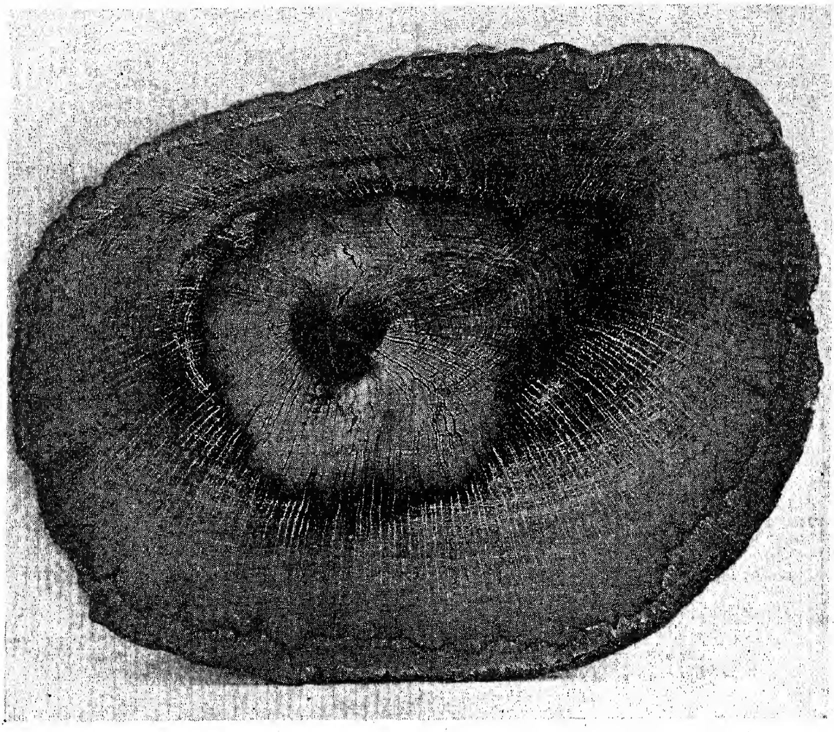

FIG. 78.—Cross section of a petrified tree trunk that is completely replaced by opal (silica and water). The structure of the original woody tissue is preserved. (*Courtesy American Museum of Natural History.*)

Ground Water as a Solvent.—Ground water continually is dissolving material from the rocks below the surface of the earth. All spring water contains dissolved mineral matter. The formation of subterranean caves and channels is evidence of this process. If ground water were chemically pure, relatively little mineral matter would be dissolved; but since most ground water is of meteoric origin, carbon dioxide, oxygen, and other gases were added to it as it fell through the atmosphere. Furthermore, decomposing organic matter in the soil near the surface adds organic acids to the percolating water, and these greatly increase its solvent power. As it penetrates to greater depths it is heated and its solvent action is increased. Since some minerals are more soluble than others, it carries away first those constituents of the rocks that are dissolved most readily and thereby makes the rock more porous.

Because of the greater resistance to solution of certain minerals extensive deposits of economic value have been concentrated at or near the surface by the solution and removal of the valueless minerals associated with them. The extensive residual iron ores of Cuba, of the Appalachian region, and of the Lake Superior region in the United States, together with great deposits of manganese and aluminum in various parts of the world, owe their concentration to the solvent action of ground water.

Replacement or Substitution.—Where solution and deposition are in progress simultaneously, one mineral may be dissolved and other mineral matter deposited in its place. The process is *replacement.* If the material replaced is of organic origin, it becomes *petrified* by the mineral matter that replaces it. Thus if a log or stump is buried in a bed of sand or volcanic ash that later becomes saturated with ground water, the replacement of the wood by silica is accomplished slowly as the wood decays (Fig. 78). Eventually, a large tree trunk may be converted into a solid mass of silica. The famous petrified logs and forests of Arizona owe their origin to this process. Erosion has again exposed them, so that, at present, silicified stumps and logs occur at the earth's surface (Fig. 79).

FIG. 79.—Petrified tree stumps, Yellowstone National Park. The tree trunks are no longer wood but are replaced by mineral matter.

Deposition by Ground Water.—A considerable portion of the mineral matter taken into solution by ground water is again deposited before it is carried far. Subsurface waters in large measure are acidic solvents in the zone of leaching above the ground-water table, but beneath this surface the water moves so slowly that it remains in contact with the rock walls of the fractures and pores for so long a time that its acidity is neutralized and precipitation ensues. The precipitated minerals tend to fill the spaces and to cement the rocks more thoroughly for some distance below the ground-water level. The zone throughout which such deposition takes place is known as the belt, or *zone of cementation.* At places mineral-bearing waters deposit their loads upon the walls of cracks or joints to form veins. This type of deposition takes place also when hot, magmatic waters or vapors penetrate the rocks. The precipitation is brought about by the cooling effect produced by the fissure walls, by surface waters, or by chemical reactions with the minerals lining the fracture. Thus an acid solution flowing through a fissure in limestone would become neutralized in contact with the calcium carbonate, and the mineral matter in solution may be deposited to fill the fissure and thus form a mineral vein.

Under certain conditions chemical precipitation takes place about some nucleus, such as a leaf or animal remains or even around a pebble. Deposition once started seems to lead to further precipitation on the same surface, and concentric layers of mineral matter are deposited. Rounded, irregular bodies called "concretions" are formed in this manner (Fig. 291). They are most commonly formed in rather porous sedimentary rocks. Many coal beds contain concretionary masses of iron sulphide in the form of the mineral marcasite. At some places small nodular concretions of calcium phosphate have formed in sufficiently great numbers to be of commercial value as a chemical fertilizer.

Underground Channels and Caves.—Ground water sinks readily in regions where there are cracks or joints in the rocks. These are gradually enlarged by the corrosive work of the descending water and at certain places become greatly extended parallel to the bedding where the rock has layers that are easily dissolved (Fig. 80). In time these channels become large so that considerable surface water drains into them. Where the openings to the channels form conspicuous holes at the surface of the earth, they are termed *sink holes,* or swallow holes. Such sinks are most commonly formed in limestone, gypsum, and salt. In other rocks the insoluble parts left as a residue are far larger than the percentage dissolved. In such rocks, therefore, the soil or rock waste fills the space originally occupied by the rock, and no such cavities result. Where pure limestone is attacked, the whole rock is soluble, and therefore nothing remains to fill the spaces where percolating waters dissolve the strata. Thus in the course of time an elaborate system of spacious tunnels and chambers may be dissolved out of solid rock. The vast labyrinth of the Mammoth Cave of Kentucky and the intricate grottoes of Carlsbad Cave in New Mexico, Luray Cave in Virginia, Wyandotte Cave in Indiana, and many other famous caves on this and other continents owe their origin to the solvent action of ground water.

In regions where caves and sink holes abound, some of the roofs of the caverns collapse, and a very irregular type of topography is developed. The slopes to the sink holes and also the slopes of the elevations between

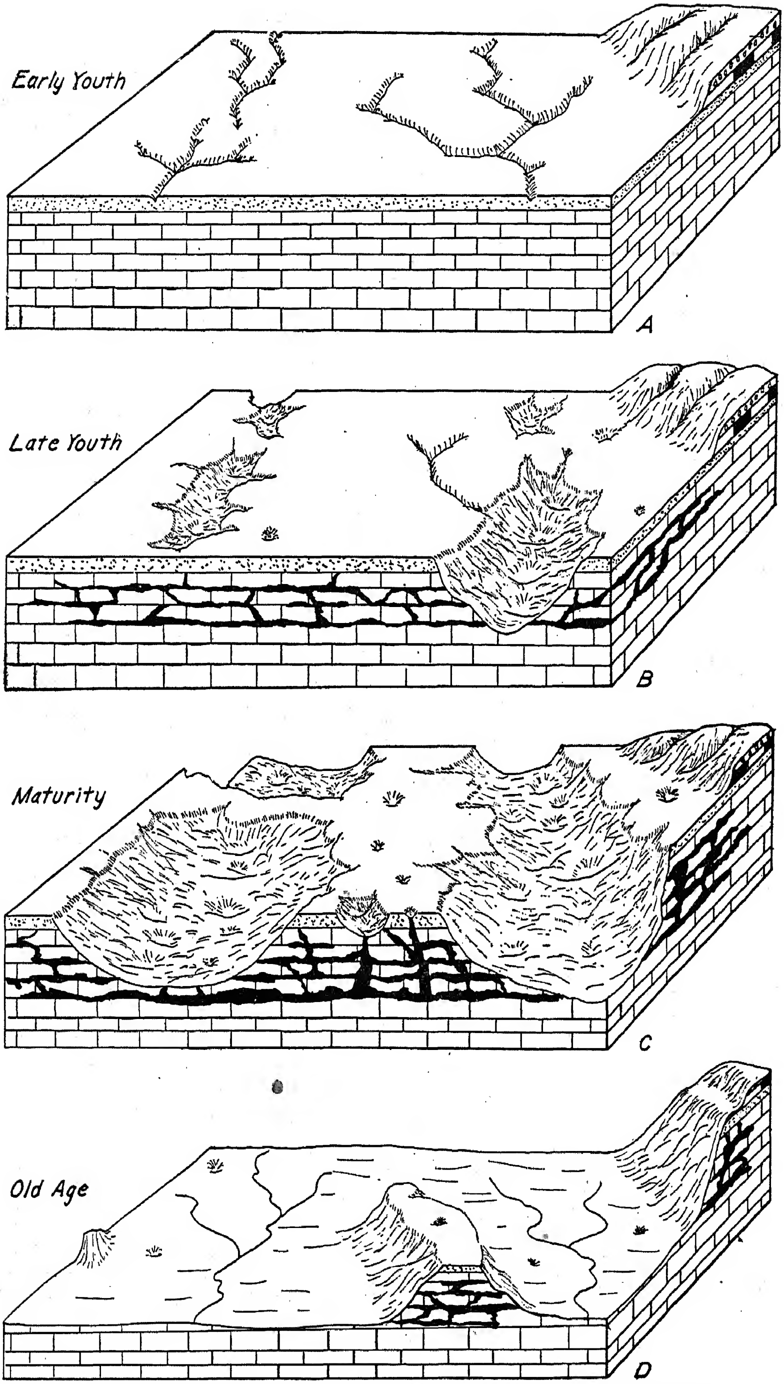

Fig. 80.—Four stages of erosion of an area with flat-lying limestone in which under ground drainage is developed. The caves are shown black. (*After A. K. Lobeck, Kentucky Geol. Survey.*)

them become steep and cliff-like. The surface is etched out in a network of numerous short gullies and ravines, which terminate abruptly where they discharge their waters into subterranean channels, and consequently the surface is rough. Such topography is characteristically developed in the Karst Mountains, northeast of the head of the Adriatic Sea in an area composed of limestone. In the United States similar topography is developed in limestone areas in central Tennessee and Kentucky and is referred to as *Karst topography.* Occasionally the bottoms of sink holes become choked with débris, and small lakes or ponds are formed. Alachua Lake, Florida, is an example. Prior to 1871 the surface drainage of Alachua prairie emptied into a large sink hole. That year the outlet of the sink was clogged, and a lake nearly 8 miles long and 4 miles wide

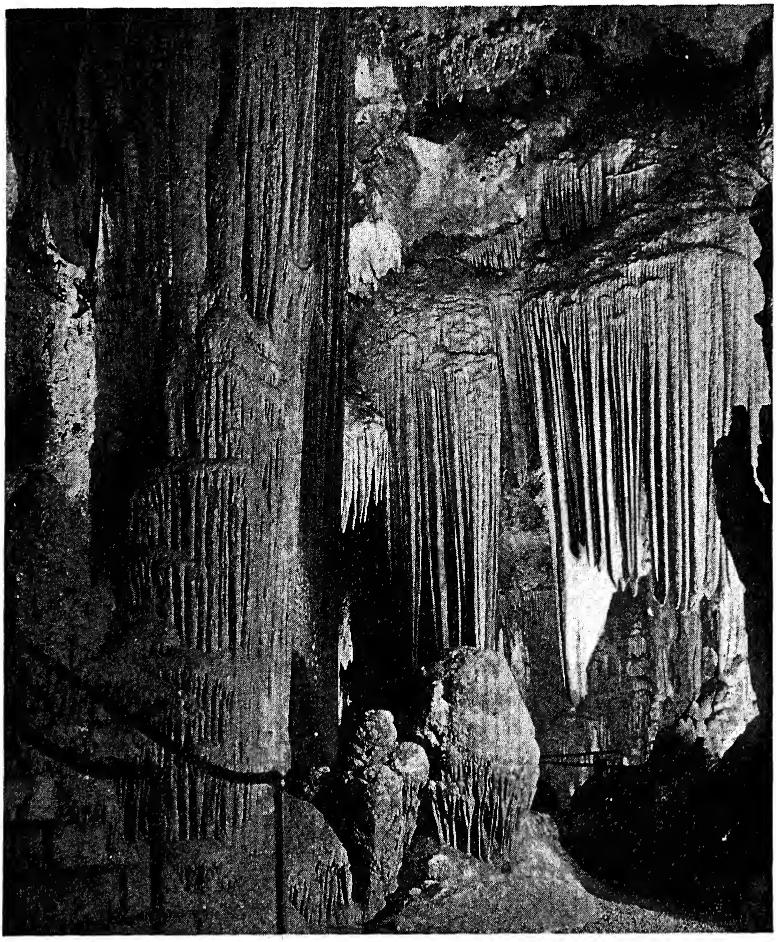

Fig. 81.—Stalactites, stalagmites, and pillars in the caverns of Luray, Virginia. (*Courtesy Luray Caverns Corp.*)

was formed. About 20 years later the outlet was again opened, and the lake was drained underground.

Deposits in Caves.—Most of the mineral matter deposited in caves is calcium carbonate. It assumes various forms, among them the *stalactites*, which are attached to the roof of the cavern or to some projecting edge, and the *stalagmites*, which form on the floor of the cavern and build upward, forming mounds and cones on the limestone floor (Figs. 81, 82).

Stalactites assume many shapes, determined by the manner in which the water trickles over them and by the amount of water present. Beautiful forms fringed with crystals of calcite, curtain-like draperies hanging from the roofs, grotesque shapes that rise from the floors, and pillars ornamented with many varieties of sculpture may be observed in the same cavern. A stalactite broken across shows a radial structure, with fibrous crystals passing across concentric zones of growth. The growing stalactite is kept moist by calcium-bearing water trickling over its surface and is lengthened and extended from the center of the structure. Stalactites have their beginning on the damp roof of the cave where drops of water gather and begin to evaporate and thus lose carbon dioxide. The drops then become saturated with carbonate and deposit the excess as a ring at their margins. Drop after drop lengthens the ring into a long pendant which later becomes a solid stalk. Those that reach the floor of the cave may thicken into massive pillars. Many pillars are formed also by the union of stalagmites that grow upward from the

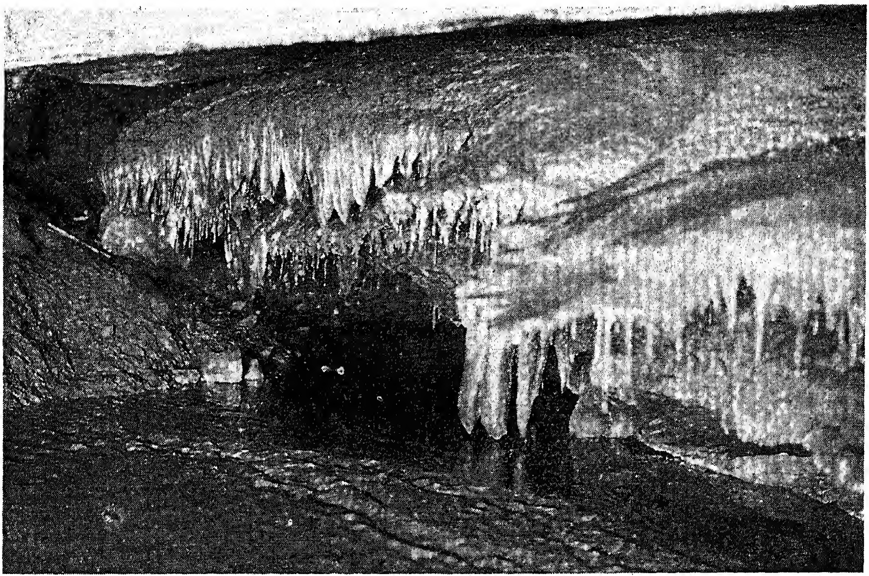

FIG. 82.—Underground stream in Endless Caverns near New Market, Virginia. (*Courtesy Endless Caverns Corp.*)

floor with stalactites that hang from the roof. Some of the deposits in Carlsbad Cavern, New Mexico, are extensive; Giant Dome, for example, is an enormous stalagmite, about 16 feet in diameter and 62 feet high. Twin Domes is another stalagmite mass more than 200 feet wide at the base and over 100 feet high.

Famous Caves.—One of the largest caves in the world is the Carlsbad Cave in New Mexico, located in a region of limestone rocks in the Guadalupe Mountains. Because of its fantastic display of ornamental stalactites, stalagmites, pillars, curtains, and frescoes of onyx, it has recently been proclaimed a national monument and added to our national park system. Perhaps the best known of American caverns is Mammoth Cave in Kentucky. Some of its caverns have been studied and mapped in detail, but there are many others that have never been fully explored.

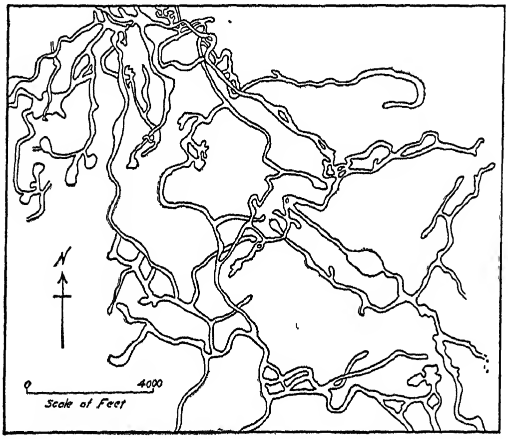

FIG. 83.—Sketch showing some of the subsurface channels in Mammoth Cave, Kentucky. (*After E. O. Hovey.*)

There are several hundred miles of connected galleries with lakes, rivers, and waterfalls in the Mammoth Cave system (Fig. 83). These galleries vary in height from a foot or two to more than 100 feet. In some parts of the cave one gallery is located above another. Mammoth Dome, which is an expanded portion of the cavern, is about 400 feet long, 150 feet wide, and from 80 to 250 feet high.

Luray Cavern in the Shenandoah valley, Virginia (Fig. 81), is famous for its brilliantly colored stalactites, of which there are as many as 40,000 visible from a single point. One celebrated group is that of the Swords of the Titans composed of eight staffs, 50 feet long, 3 to 8 feet wide, and as much as 2 feet thick. Other noted limestone caverns in the United States are the Wyandotte and Marengo caves in Indiana, Wind Cave in South Dakota, and Marble Cave in Missouri. One of the best known caverns is at Adelsberg in Italy. Its four great chambers are visited frequently, and festivals are conducted in its grottoes.

Mechanical Work of Ground Water.—Underground waters move too slowly to accomplish much mechanical erosion. Locally, however, caves may be formed by water moving along joints in a poorly cemented sandstone. The ground water dissolves the cement and carries out the grains of sand. Thus Carver's Cave was formed in the St. Peter sandstone at St. Paul, Minnesota. Ground water derived from prolonged rains or from melting snow may saturate masses of talus or other rock débris resting on a sloping surface, until its weight is increased and its traction reduced to a point where gravity causes it to glide downward as a landslide. Generally a soft, clayey rock surface well lubricated by ground water is necessary to initiate sliding. Under such conditions the water which percolates down through the porous material reaches the impermeable clay, which becomes soft and slippery, and the overlying mass slides over it to a lower level. In mountainous regions a series of landslides may give rise to a large train of rock débris that continues to move gradually down the mountain valley. Such creeping masses are referred to as rock streams or rock glaciers. In high mountains ice may accelerate the movement.

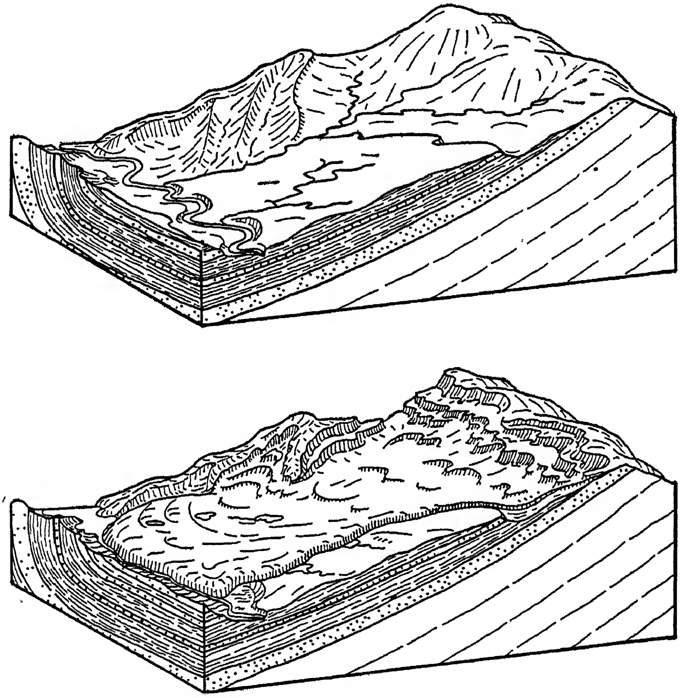

FIG. 84.—The Gros Ventre landslide, northwestern Wyoming. (*Redrawn from Blackwelder.*)

Many destructive landslides have taken place during the past few centuries. In 1855 a mass of rock débris 3,500 feet long, 1,000 feet wide, and 600 feet high descended into the valley of the Tiber River. It formed a dam across the valley so that the village of San Stefano was flooded to a depth of 50 feet. Many lives were lost and buildings were destroyed. Along the Lievre River, north of Buckingham, Quebec, an area of about 100 acres slid into the river. A clay terrace was resting on gneissic rock, and after several days of rain the clay had become saturated with water and slid under the additional weight. The momentum developed was so great that large masses of the clay were thrust up the opposite bank of the stream to a height of 25 feet. Another destructive slide occurred in Canada in 1903 at Frank, Alberta, where the entire face of Turtle Mountain, estimated at 40,000,000 cubic yards, broke loose and was dashed to the base of the mountain and hurled across a valley and 400 feet up the opposite side. The length of the slide was about 2½ miles. The entire period of movement was less than 2 minutes. In the Columbia River Gorge near Stevenson, Washington, numerous slides are continually creeping toward the valley.

In the Gros Ventre valley, Wyoming, south of Yellowstone Park, a landslide (Fig. 84) continued through the years 1908 to 1911, obstructing the river and wagon road and causing the formation of small lakes and ponds. The movement was slow like that of a glacier and was due to the slipping downward on a rather steep slope of rocks underlain and lubricated by soft unctuous clays.[1]

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CHAPTER V

GRADATIONAL WORK OF STREAMS

Introduction.—The gradational work of streams includes degradation or wearing down of rocks by streams and aggradation by deposition of the degraded material. Water falls upon the earth as rain, sleet, hail, and snow. Some of it is evaporated, some soaks into the ground, and a third portion is carried off by the streams and finds its way to the sea. The total annual rainfall on all the land areas of the earth has been estimated to be about 26,679 cubic miles. About 22 per cent of the precipitation is carried off by streams and constitutes the *run-off*. This run-off is the chief degrading agent on the land surface and is, therefore, the most important factor in the process of erosion (Fig. 85), or the reduction of land areas to lower levels. Since the average elevation of the con-

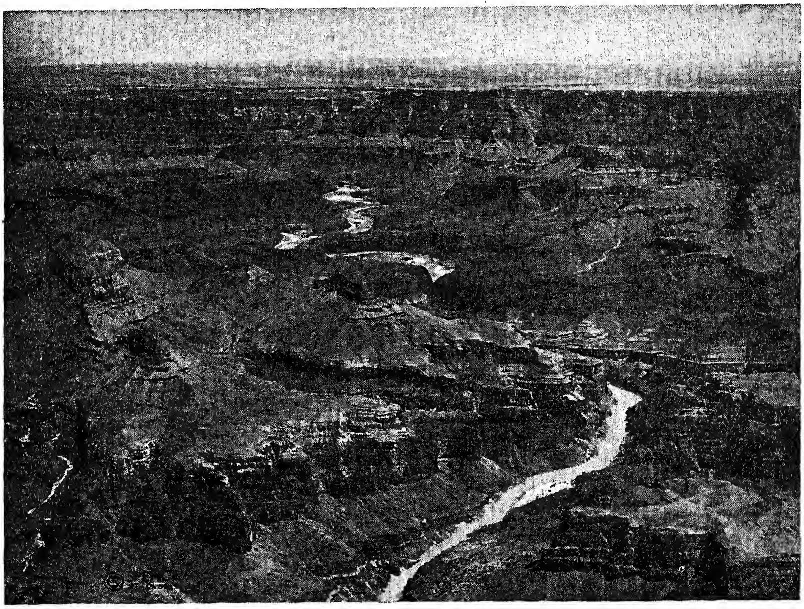

Fig. 85.—The Grand Canyon of the Colorado River looking east toward the Painted Desert. (*Photograph by Spence Air Photos.*)

tinents is about ½ mile and the amount of water carried by streams is about 5,800 cubic miles per year, the energy available for eroding land areas is great.

Distribution of Rainfall.—The work of running water begins with the raindrop. Drops run together to form rills, and these join to form brooks, creeks, and rivers; there are no sharp dividing lines between them. The amount of rainfall which a given area receives has a direct bearing on the rate at which erosion will proceed. The distribution of rainfall is influenced by the winds, which carry water vapor from places where it evaporated to regions where the temperature favors its condensation and precipitation. Therefore to understand the rainfall of any given region, it is necessary to know (1) the prevailing and periodical winds, (2) the topography of the surface over which the winds have previously passed, and (3) the physiographic features of the region itself.

The amount of rainfall is very unequally distributed over the earth as a whole. At certain places, as in parts of India, it is over 500 inches a year, whereas in the Sahara desert it is less than 10 inches. On the Atlantic coastal plain of North America it is approximately 40 inches a year. Farther west in the northern portion of the interior basin it is 30 inches, and on the Great Plains it is 20 inches or less. In the Great Basin province of the southwest, between the southern Rockies and the Sierras, in the region known as the American desert, less than 10 inches of moisture falls per year.

Run-off.—That portion of the rainfall which flows off the surface of the land in the form of visible streams is the run-off. Its amount is not uniform, even in areas with equal amounts of precipitation. In regions of marked relief the run-off will exceed that of the plains or prairies, and unconsolidated sediments or soils will absorb more rain than indurated rocks. Porous formations, however, when saturated with water allow a high percentage of water to run off. The character and amount of vegetation also influence the run-off. A heavy sod on a slope may shed water like a thatched roof. Most vegetation, however, holds water and delays the run-off. This is true especially of decaying plant tissue, for it acts like a sponge and absorbs large quantities of water. Weeds, brush, and logs also delay run-off by damming the surface water that otherwise would be free to flow away. The removal of forests, therefore, tends to increase the variability of the streams' volume and to increase the size and frequency of floods.

Dry, hot winds evaporate much of the water which remains on or near the surface. In arid regions where the humidity is low the winds are so dry that evaporation goes on rapidly. In the moist tropics, however, the humidity is high, and exposed surfaces remain damp. Under such conditions a smaller part of the rainfall is evaporated.

Ratio of Run-off to Rainfall.—The percentage of the rainfall discharged by rivers cannot be measured precisely, for no rule can be made to apply to all parts of a continent. In general, in areas with an annual rainfall of 50 inches about one-half is discharged by rivers; whereas with a rainfall of 20 inches only 15 per cent is discharged. The Ohio River discharges about 30 per cent of the rainfall of its basin, while the Missouri River carries away only 15 per cent. In southwestern North America, a number of the streams do not discharge more than 5 per cent of the rainfall.

DEGRADATIONAL WORK OF STREAMS

Stream Erosion.—It has been stated that water which falls on the land and is not evaporated aids in transporting the rocks of the land to the sea. This erosive work of water is accomplished by a number of subprocesses which operate jointly. These are:

FIG. 86.—Stream sediments on the bottom of a valley in Nevada. The rock fragments aid in abrading the valley. (*Photograph by Erdmann, U. S. Geol. Survey.*)

1. Corrasion, or the mechanical wear of the bed due to the friction of silt, sand, gravel, and boulders carried by the stream (Fig. 86) and to undercutting of stream banks. Abrasion, or wearing of the rock fragments in transit, commonly accompanies this process.
2. Corrosion, or the solvent action of water on the rock minerals.
3. The impact of water thrown against loose débris or into concavities and recesses.
4. Transportation, or the removal of the products of rock weathering and corrasion.

Corrasion.—The corrasive or abrasive action of clear water is slight. This is well shown in clear streams like the Niagara River, a stream of clear water from which most of the sediment has settled in Lake Erie. Delicate plants such as algae grow at the very brink of the falls and form a green coating on the rocks. The current is very swift, but the force of the torrent is unable to tear the tiny plants from the rocky bed. But when the running water transports grains or pebbles of mineral matter, it becomes a powerful agent of erosion, and the results of its work are seen in the excavation of deep canyons and gorges. The work of a stream is accomplished chiefly by means of the bottom load or by sand, pebbles, and silt which it sweeps along near the bottom of the stream. With these as tools it grinds and rasps the rocks of its bed as sandpaper or as a file abrades.

The corrasive power of river water varies as the square of the velocity of the stream. This relation is appreciated when it is recalled that if the speed of the current is doubled, it will hurl twice as many sand grains as before in the same period, and it will throw each grain against an exposed rock in the stream bed with twice the force at its former speed. Thus the rock surface will be eroded four times as fast as it was before the speed was doubled. There are other factors also to be considered, such as the character and amount of the transported material and the character and structure of the rocks through which the channel is excavated.

Corrosion.—Chemically pure water does not exist under natural conditions. Because of its solvent power, it is ever charging itself with impurities, many of which greatly increase its efficiency as a solvent. The water of many streams, especially after flowing through bogs and marshes where decaying vegetation abounds, is charged with carbonic acid in solution. With the aid of this acid, together with atmospheric oxygen, stream water acts on the rock surfaces with which it comes in contact. The rate of dissolution of the rocks is almost imperceptible, except where calcareous sediments are corroded along joints and fissures. The amount taken in solution by any one stream may seem small, but in the aggregate, the amount of material thus dissolved from the land and carried into the sea is great. It is estimated that about 5,000,000,000 metric tons of solid material go into solution on the continents annually, but the greater part of this is contributed to the streams by ground water.

TRANSPORTATION

Sources of Materials.—The materials carried by a stream are its load. The load is derived from a number of sources: (1) the larger part is supplied by the weathering and removal of rock from the slopes of its tributaries. During a rain the immediate run-off is muddy with waste as it rushes along gullies or washes down the hillsides. This is true of cultivated regions, especially where plowed fields lie on the slopes and numerous rills and minor tributaries carry the unconsolidated material to the larger streams. (2) Some of the load that a stream carries is obtained by wearing it from its banks or bed, and (3) some of it may fall into the river from steep banks where it has been dislodged by the pull of gravity upon weathered débris. (4) In regions with sparse vegetation, earth particles are moved by the wind, and sand or dust may be dropped into the stream to increase its load. (5) Great numbers of streams that owe their origin to the melting of glacial ice are turbid and loaded with silt. The ice and water from its melting carry "rock flour" produced by the grinding of the boulders held in the ice. (6) In regions of volcanic activity vast quantities of dust and ash are discharged into the atmosphere, and some of it falls into streams or is carried into them by rills during subsequent rainfall. During the recent eruption in the Katmai region of Alaska some streams were completely clogged with ash, and small boats were filled and buried under the débris. (7) Minor methods by which streams are supplied with material include the impact of driftwood or of floating blocks of ice on the walls of the stream channels, disturbances produced by the uprooting of trees, and those produced by the work of animals and plants.

Transportation of Load.—A body immersed in water loses weight equal to that of the water displaced. Most of the mineral and rock fragments carried by a stream weigh about two and one-half times as much as water. Immersed in water they therefore lose two-fifths of their weight. Water films, moreover, are attached to small particles, rendering these lighter and facilitating their transportation by running water.

Streams move their loads (1) by pushing and dragging many of the angular pieces; (2) by rolling the rounded and subangular pebbles along their floors; (3) by carrying in suspension the fine grains of sand, clay, and silt; and (4) by dissolving and carrying in solution the more soluble compounds. Because of the irregularities of the stream bed, the velocity of the current varies at different places along the stream; and since there is greater energy where the velocity is increased, the movement of the sediment is not uniform. The eddies and cross currents produced by the deflection of the water from the irregularities of the stream bed tend to keep particles in suspension which would otherwise settle to the bottom of the stream. Many particles fall to the bottom many times during their journey and remain lodged on the floor of the channel until they are lifted up by deflection currents to near the surface where the velocity is greater. Most of the material carried a short distance above the floor of the stream proceeds by a series of short leaps or jumps. This type of progress is "saltation." *Stream traction* is that process by which rock material is forced downstream by pushing, rolling, and saltation.

Relation of Velocity and Transportation.—The velocity of a stream is determined by (1) the slope or gradient of the stream bed, (2) the shape and configuration of the valley walls, (3) the volume of water in the stream, and (4) the amount of sediment the stream is carrying.

Since the flow of water is due to gravity, it is obvious that in general the greater the fall per mile the faster the water will flow. The slope commonly decreases from headwaters toward the mouth of a stream, and consequently the velocity diminishes also. The average slope of the large rivers of all continents is approximately 2 feet per mile. For many of the navigable streams it is less than 1 foot per mile. Irregularities of the channel walls and bottom cause the water to be checked by friction, and it follows that the smoother and narrower a channel the lower will be the loss of energy because of friction. A stream bed studded with boulders or one running at right angles to rough rocky ridges on its floor has obstacles that check the velocity of the stream.

An increase in volume accelerates the rate of flow of a stream by bringing about increased depth of water without greatly increasing the amount of friction. The rate of flow varies from time to time as the source of supply of its water varies. In many regions this variation is a periodic one, as, for example, rivers whose sources are in snow-covered mountains, like the headwaters of the Colorado River, where the flow is faster when the volume is increased by the melting of the snow in the spring.

The following table shows the velocities of bottom currents in a river that are necessary for the currents to move materials of different sizes.

Nature of Material Carried	Velocity of Current, Meters per Second
Fine sand	0.28
Coarse sand	0.40
Gravel	1.62
Boulders	11.69

Along valleys with steep slopes where heavy rainfall soaks the loose products of rock waste that have accumulated on a clayey surface or on consolidated rock, landslides frequently take place, and these may form temporary dams across river channels. When the ponded water overflows or breaks through the dam, it carries everything before it. In this way even small brooks become powerful erosive agents and displace hundreds of tons of material from the walls of their valleys in a few hours.

Amount of Load.—The quantity of material transported by a river is not constant, owing to the varying volume and accompanying varying velocity and also to the variable amounts of rock waste supplied by tributaries. In order to ascertain the total load, the average annual discharge of water must be determined, and the average amount of solid sediments transported and of salts carried in solution. The amount of water discharged may be ascertained by multiplying the number of square feet in the average cross section of the stream by its velocity per second, giving the discharge per second in cubic feet. The amount of silt to a cubic foot of water is found by filtering samples of the water taken from different parts of the stream during different seasons. The amount of salts in solution is obtained by evaporating filtered samples, and the composition of the salts is found by chemical analyses.

The Mississippi River annually carries to the Gulf of Mexico about 22,000,000,000,000 cubic feet of water, containing in suspension 340,500,000 tons, rolling on the bottom 40,000,000 tons, and carrying in solution 136,400,000 tons or a total load of 516,900,000 tons of rock waste. According to recent estimates more than 800,000,000 tons of material in all are carried by the rivers of the United States each year. Miller states that a train of ordinary freight cars long enough to carry this load would reach around the earth six times in the region of the equator. The total load carried in solution per year by all the rivers is the equivalent of about 100 tons for every square mile of land surface on all of the continents.

Abrasion of Load in Transit.—Dragging and rolling rock fragments over one another and over the surface of the stream bed results in wearing of the transported pebbles themselves. Such wearing is *abrasion*, or *attrition*. Because of this action, the load is finer near the mouth of large rivers, and as one goes upstream an increase in the size of rock fragments carried by streams is noted. A further reason for such distribution is that the velocity of the stream decreases toward the mouth, and hence the stream is unable to move coarse material. A high percentage of the fine silt in the lower courses of a stream, however, is derived from the attrition of the coarser material gathered by its headwaters. The effect of attrition is well shown in the table below from observations on the Mur River in Germany.

Rate of Denudation.—A river system with its numerous tributaries covers the land with a network of watercourses, which carry their loads

Average Size of Fragments at and below the City of Graz, on the Mur River in Germany

Distance Carried	Average Size of Fragments, Cubic Centimeters
At the city of Graz	224
5.68 miles below city	184
14.76 miles below city	132
24.42 miles below city	117
31.81 miles below city	81
40 33 miles below city	60
57.36 miles below city	33
68.16 miles below city	21

toward the trunk stream. With a uniform amount of rainfall the rate of denudation will be greatest in the region of the headwaters of the stream, somewhat less over the more gentle slopes of the intermediate zone, and least in the level areas near the coast line. In a large drainage basin the rate of stream erosion is influenced by many factors such as (1) the velocity of the stream, (2) its volume of water, (3) the nature and amount of its load, and (4) the character of the rocks or soils over which it flows.

In the headwaters the gradient of the stream bed is steeper, the velocity is greater, and the transporting power is higher than near its mouth. If the fragments collected by the headwaters of a river are composed mainly of resistant minerals such as quartz, while the floors and walls of the valley farther down its course are composed of softer rocks. the valley will be eroded much more rapidly than if the load of the stream is mostly clay or silt. To cut rapidly the stream must carry some resistant sediment as tools but not so much as to decrease its velocity, for then the force of its tools is diminished. Sedimentary rocks, especially those cemented with calcium carbonate, are much more easily eroded than massive igneous rocks. Thin-bedded sediments or the presence of joints and fissures favor rapid erosion.

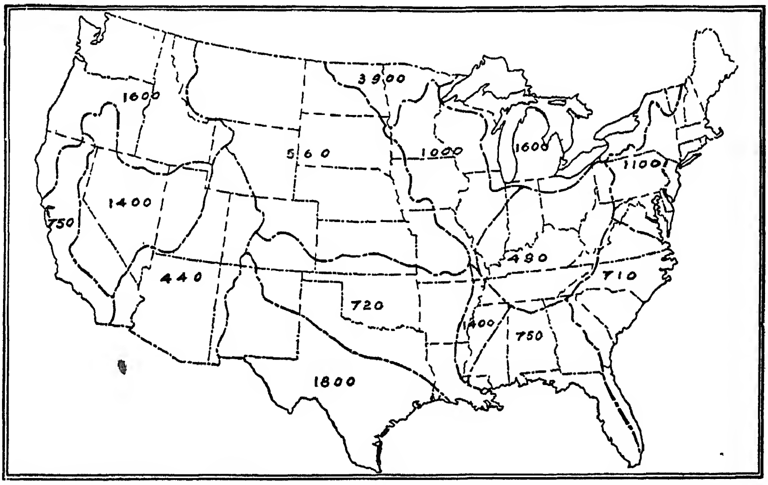

Fig. 87.—Outline map of the United States showing the estimated number of years required for the land to be lowered 1 inch by erosion. (*Data from National Conservation Commission.*)

Estimates of the rate at which certain rivers are eroding the areas that they drain have been made by various groups of investigators (Fig. 87). The Mississippi River and its tributaries, draining an area of approximately 1,265,000 square miles, is now lowering its basin at the average rate of about 1 foot in 9,000 years.

Soil Erosion.—Soil is the layer of disintegrated mantle rock formed by weathering. Generally it is colored dark by decaying organic matter, and it contains the many substances that promote plant growth, these being derived from the rocks, from the air, and from water.[1] Each year parts of the soil are washed away, and each year additions are made by weathering (page 51). If the additions are less than the losses, the land deteriorates.

Where soils are cultivated and there is no protective cover, water and wind erosion remove the soil more rapidly; and if no protective measures are provided, the soil is washed away. Sheet erosion removes great quantities of the soil, carrying it away in rills, creeks, and rivers. Water is concentrated in gullies which cut up the land and ultimately destroy it (Fig. 88). In the United States 50,000,000 acres of once productive land already has been essentially ruined for practical crop use, and as much more is in nearly as poor a state.

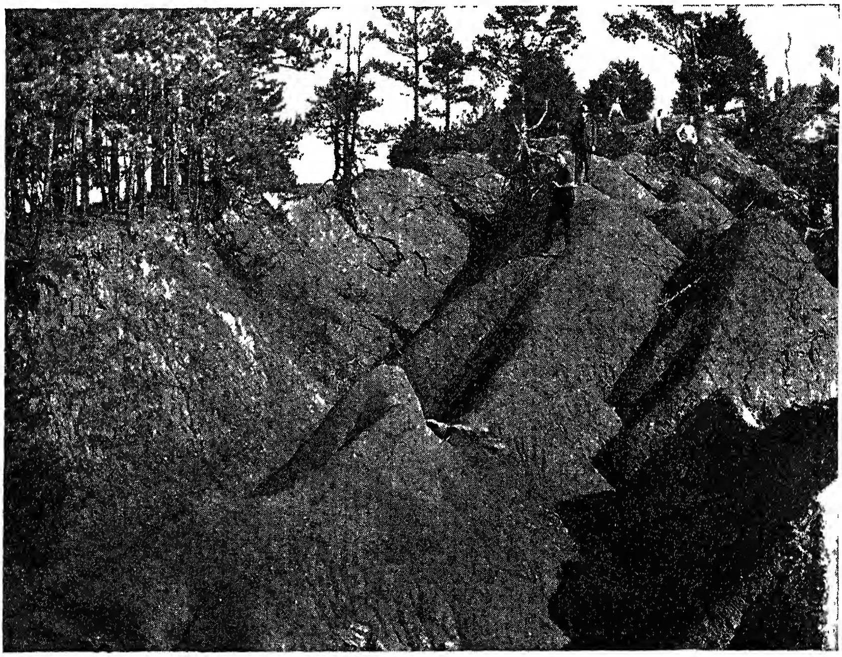

FIG. 88.—Before planting—gullies on an upper Tennessee Valley farm. Photographed December 5, 1933, just before treatment. (*Courtesy Tennessee Valley Authority.*)

The most severe erosion takes place on slopes and particularly on slopes where cleanly tilled crops such as cotton, corn, and tobacco are grown. Soil erosion is much less severe where the grasses, alfalfa, and the small grains are grown. Forests protect the soil from erosion, and

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[1] BENNETT, H. H., Soil Erosion and Its Prevention, in Our Natural Resources and Their Conservation, pp. 65–100, John Wiley & Sons, Inc., 1936.

where these are cut down and particularly where the slash and undergrowth are removed or burned, erosion is greatly increased.

Experiments at Bethany, Missouri, with a mean precipitation of 33.5 inches, on an 8 per cent slope that was cropped continuously to corn showed a loss of 67.4 tons of soil per year per acre. Where corn was rotated with wheat and clover, less than half that was lost; and where alfalfa was grown, the loss under the same conditions was only 0.2 of a ton of soil per acre. The water lost by run-off was greatly increased as soil loss increased; it was 26 per cent where corn was grown and only 3.4 per cent where alfalfa was grown. Grass crops are almost as effective as alfalfa in conserving soil and decreasing run-off.

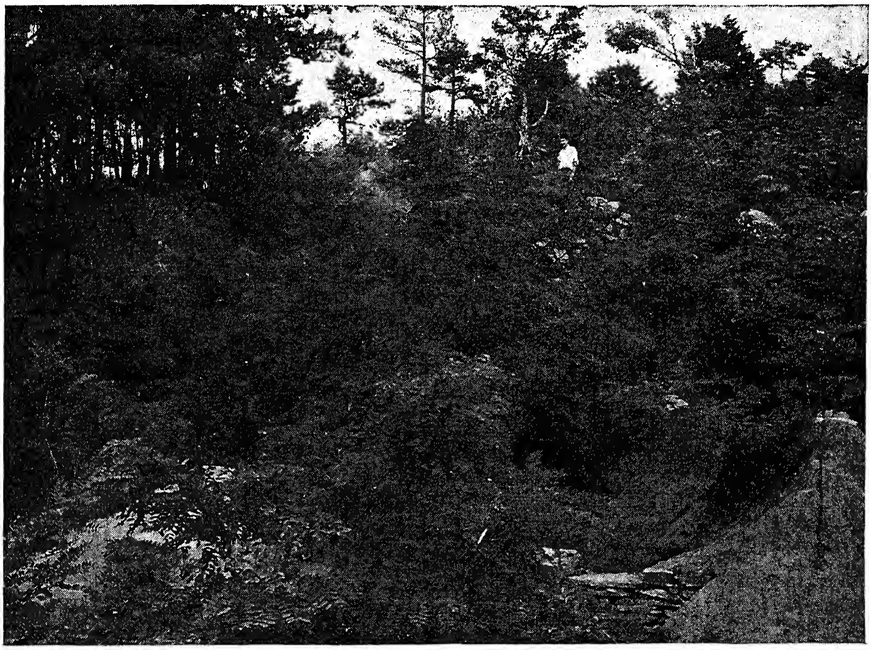


Fig. 89.—After planting—same gullies as shown in Fig. 88 on upper Tennessee Valley farm after treatment. Photograph taken July 24, 1935. Treatment consisted chiefly of planting of black locust seedlings. Some stone check dams were built. (*Courtesy Tennessee Valley Authority.*)

Various methods are in use to decrease soil erosion and land destruction. Dams of earth, rock, or logs are built in gullies to check their growth, and vegetation is planted on the bare surfaces to hold the soils (Fig. 89). Where slopes are farmed, strips of tilled crops are planted along contours to alternate with strips of small grain, grass, or other nontilled crops (Fig. 90). These catch the rain water flowing down the slopes, spread it out, and protect the tilled ground from erosion. They greatly increase the water absorbed by the ground and decrease the run-off.

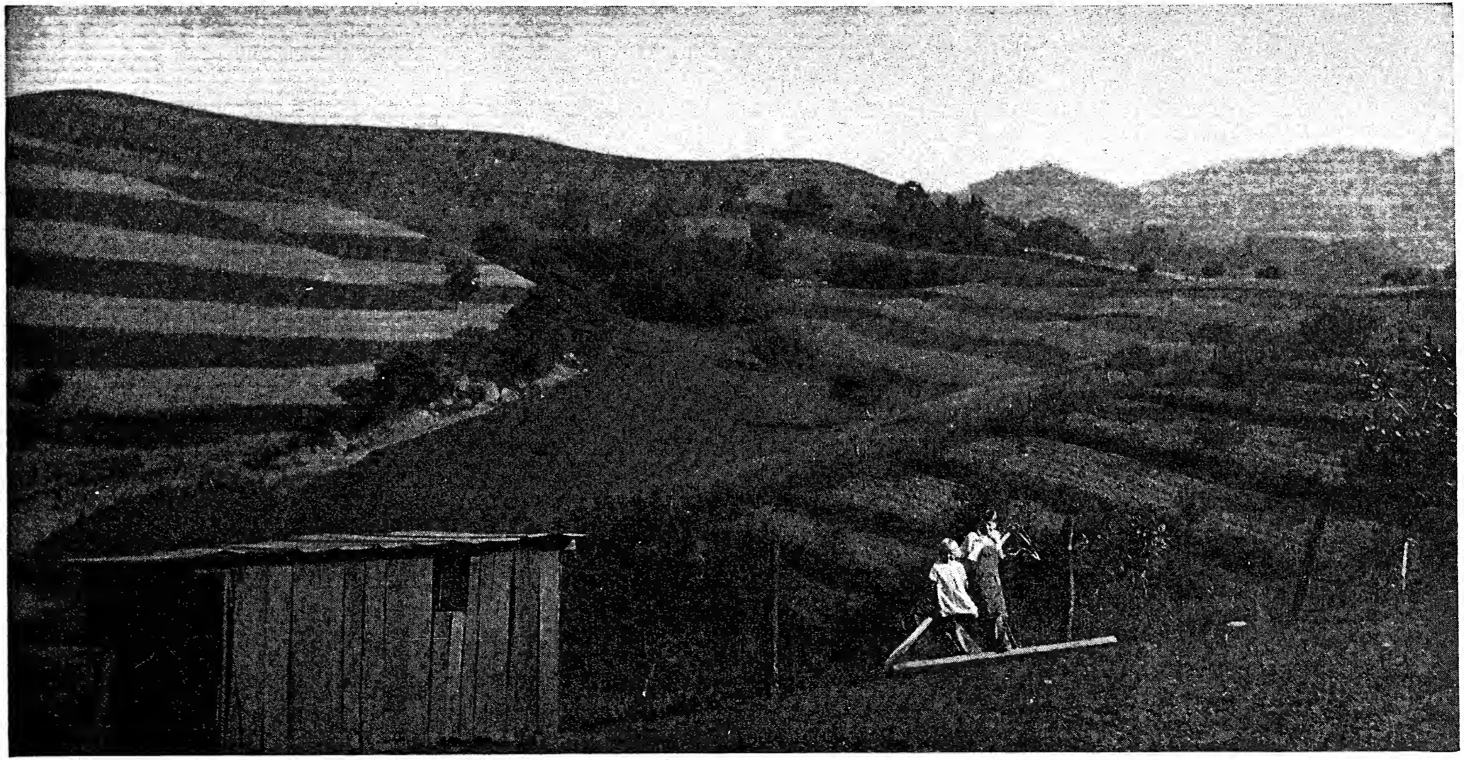


FIG. 90.—View of Zimmerman farm, Madison County, North Carolina, showing strip cropping developed by Mr. Zimmerman over a period of years to control erosion, hold water in the soil, and maintain fertility. Most of this land is kept in sod the year around, and when cultivated it yields several times more abundantly than the neighboring poor mountain farms on the same types of soil. (*Courtesy U. S. Department of Agriculture.*)

Terracing, trenching, ridging, and furrowing along contours also are at places effective. These are preventive measures. Where lands already are deeply trenched with many gullies, the problems are those of reclamation. Small dams and terraces are built to catch the soil in transit, and small rich garden and orchard plots thus are provided. Certain areas have been reforested, and others put to grass. With all of these methods of conservation, land destruction by erosion (Fig. 91) steadily gains in the United States where rainstorms are violent and tilled crops are cultivated over vast areas.

FIG. 91.—Aerial photograph of fields gullied by headward erosion. The lined areas are cultivated fields which are being reduced in size by the gullies (dark, tree covered) that are encroaching on the fields from all sides. (*Mark Hurd Air Mapping Company.*)

DEVELOPMENT AND ALTERATION OF VALLEYS

A large part of the water that falls as rain runs off the surface of the region where it falls. If the surface has a smooth, uniform slope like a gently pitching roof, it flows off as a sheet, and no channels are developed. Such smooth surfaces, however, are rare. Slight irregularities lead to the formation of rills which erode small furrows. The rills unite to form rivulets (Fig. 88), and the uniting rivulets form torrents which cut deep gorges and canyons which are eventually transformed into wide valleys with gentle slopes. Some of the details of this transformation are discussed on the pages that follow.

Valley Growth.—The lengthening of a valley is accomplished for the most part by head erosion. That is, many streams start at or near the places of their outlets and grow in length by cutting backward into the slopes. This process continues until the gully reaches an obstacle, such as a resistant rock formation, or until it reaches a point where lack of slope or the tributaries from the opposite side of a slope stop its progress. If the erosion on the two sides of the divide is about equal, a fixed divide is established, for even though continued rainfall may tend to lower it, its geographic position remains unchanged. Where erosion on one side is more rapid than on the other, however, the divide shifts slowly toward the side of less rapid denudation.

If, in its migration headward, a stream reaches a lake, then the lake basin becomes a part of the stream system, and the streams flowing into the lake become tributaries of the same system. Its retreat is retarded at the lake until the stream cuts downward to the level of the lake basin and drains it.

Deepening of a Valley.—The deepening of a valley accompanies its growth in length, for while cutting backward, the stream wears its bed downward. During each successive downpour the gully acquires more surface water and is washed out deeper. Side gullies develop and grow into tributaries, the tributaries also acquire side gullies, and the surface water of a large area is directed through the one main channel. It follows that the erosion of the main channel will be increased, for increased volume means increased velocity with its greater eroding power. As long as the valley is shallow, its supply of water is limited to the immediate run-off at the time of rainfall. Streams that flow only during a rainy season or during a downpour of rain are "intermittent" streams. As a valley is deepened, the floor of its channel approaches the level below which the pore spaces of the rocks are full of water, or the ground-water table, and after the bottom of a valley penetrates this level its reserve of water is sufficiently increased so that it supports a stream even during the dry seasons. Streams that have their valleys cut to the ground-water level are "permanent" streams.

In regions where the gradients are steep, the downcutting by the streams will proceed much more rapidly than where the slopes are low and the streams sluggish. In fact, many slow streams make their valleys shallower, for they deposit more than they take away. Where the gradient is such that there is approximately a balance between the amount a stream erodes and the amount it aggrades, the stream is said to be *at grade*, or it is commonly referred to as a "graded" stream. In a long valley, like the Mississippi, deepening may be in progress toward its upper end, whereas the same valley may be becoming shallower because of deposition in its lower course. Where the rate of deepening exceeds that of the erosion of the banks, deep canyons with cliff-like walls are developed (Fig. 85). The depth to which a valley may lower its bed depends on the elevation of the land in which it has been cut. High plateaus such as the Colorado Plateau or high mountains such as the Himalayan range commonly have valleys of great depth.

The depth limit of a valley is determined by the level of the body of water into which its river flows. The channel of a large river, such as the Mississippi, near its mouth may be cut somewhat lower than sea level, but in general the level of the lake or ocean is approximately the depth limit of the valley of the river that discharges into it. Only the lower end of a valley ever attains this limit, for the stream bed must have a gradient, or the river will not flow. The lowest possible level to which a stream can erode its drainage basin by mechanical wear is called *base level.*

Widening of a Valley.—The widening of a valley, or lateral erosion, goes on in conjunction with its deepening. If such were not the case, all valleys would be canyons with steep walls. In some regions, conditions for downcutting are more favorable than for lateral erosion, as along the Gunnison River in Colorado and along the upper portion of Zion Canyon, Utah, where nearly vertical walls over 2,000 feet high are only a few hundred feet apart at the brink of the canyon. Most valleys are much wider than their streams, however, although the character of the rock over which the stream flows may produce local variations in the width, so that narrow portions alternate with wider ones. If a stream crosses a tilted bed of hard, resistant rock lying between softer ones, the valley will widen faster both above and below the hard bed than it does where the stream crosses it. A "narrows" will be formed where the stream cuts across the resistant rock.

Valley widening is accomplished in many ways, the most important of which are (1) by creeping or slumping, (2) by side wash, (3) by the activities of animals and plants, (4) by glacial abrasion, (5) by undercutting, and (6) by tributaries.

1. While the slopes are steep the loose products of rock weathering creep slowly downward under the action of gravity. If clayey material becomes thoroughly saturated with water during a rainy season, large masses of it may slide or slump toward the valley floor and carry débris with it to lower levels. This process is in progress on a large scale along parts of the valley of the Columbia River, where huge landslides are moving toward the valley, carrying forests and buildings with them.

2. Rain falling on the slopes of a valley washes loose material with it; and where the valley is in unconsolidated sediments, furrows and gullies are excavated in its walls. The débris is carried toward the stream channel, and the valley is widened.

3. Many land animals visit streams to obtain water or to graze along its banks; and as they walk along the valley slopes, loose débris is dislodged. Burrowing animals bring sand and clay to the surface that is readily washed down the slopes. Roots of trees aid in the disruption of the rocks along the valley; and when the trees are overturned, more material is loosened and eventually carried downward.

4. At high altitudes where glaciers partially fill the valleys, ice widens the bottom as rapidly as the walls are eroded back, and wide, flat-bottomed valleys result (see page 178).

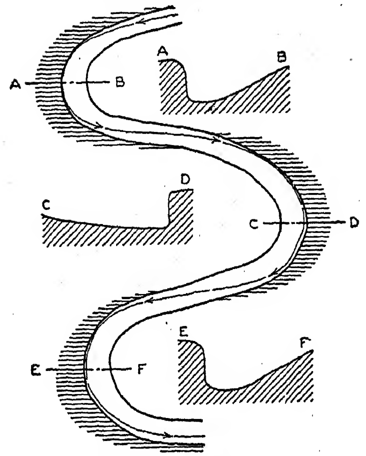


Fig. 92.—Diagram of meandering stream showing change of profile along its course. On the outer or convex side of each bend the bank is high and steep, due to undercutting, whereas on the concave side it slopes at a gentle angle.

Fig. 93.—Aerial photograph of St. Francis River in east central Minnesota, showing numerous abandoned meander loops. (*Courtesy Dr. W. S. Cooper.*)

5. The course of a stream is rarely straight, and at each curve it tends to cut more on one bank than on the other (Fig. 92). This is due to the fact that the water tends to pile up on the bend, and the outside of the curve receives a greater volume of water. In this way the velocity is increased, and the bank is undermined and moved back. As a valley approaches base level, its gradient is decreased, and the flow of its stream becomes more gentle. In such a slow stream the currents easily are deflected (1) by some obstacle on the floor of the channel or (2) by a projection of more resistant rock along its bank or (3) by the entrance of swifter currents from a tributary stream. The deflected current strikes the opposite bank and as it moves downstream tends to be thrown back again to the side from which it first issued. Thus the stream develops a series of winding curves or meanders (Fig. 93). Once started, the

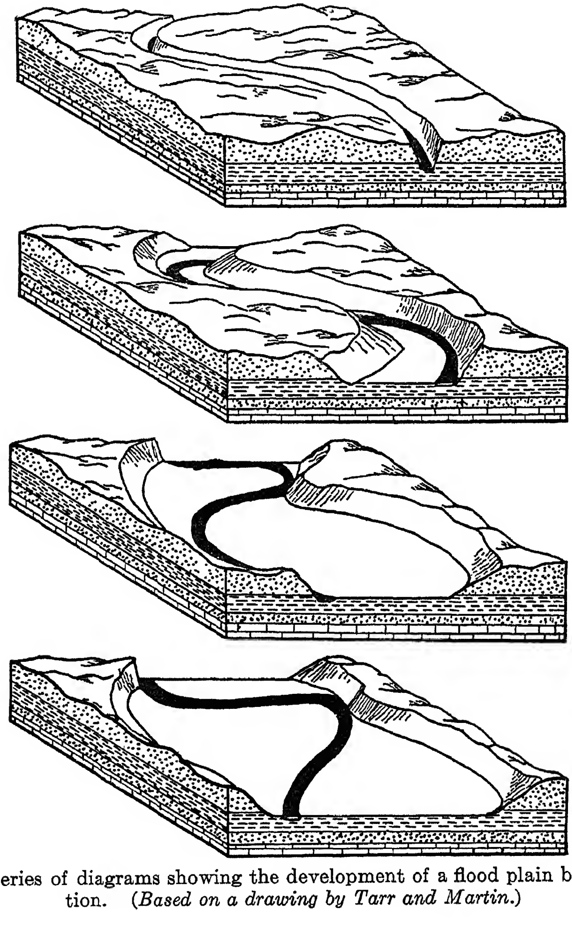


FIG. 94.—A series of diagrams showing the development of a flood plain by lateral planation. (*Based on a drawing by Tarr and Martin.*)

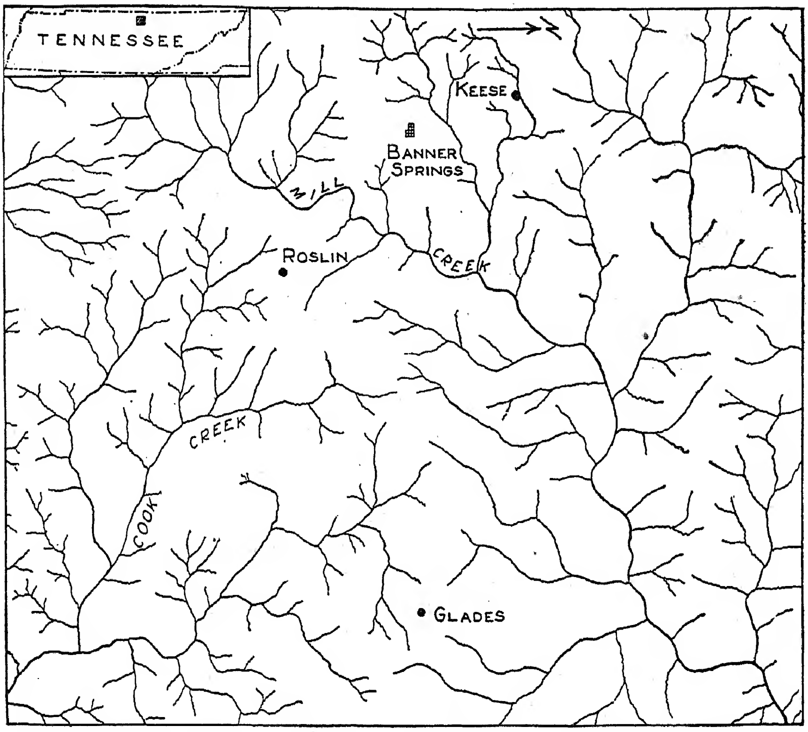
meanders become more and more pronounced, and finally a series of loops separated by narrow necks of land are developed. Eventually the stream cuts through the narrowed neck of land between two loops and leaves the meander as a long, curved *oxbow lake*. Where this happens, the river shortens its course, and the current is modified so that a new meander is likely to be formed because of the shifting of the currents. The old deserted loops with their shallow lakes are converted gradually into stagnant pools and bogs characteristic of flood plains.

Fig. 95.—Flats of the Mississippi River valley near Dubuque, Iowa, with numerous lagoons, abandoned channels, and partly filled lakes. The topography of the uplands is still in early maturity. (*Lancaster quadrangle, Wisconsin, Iowa, Illinois, U. S. Geol. Survey.*)

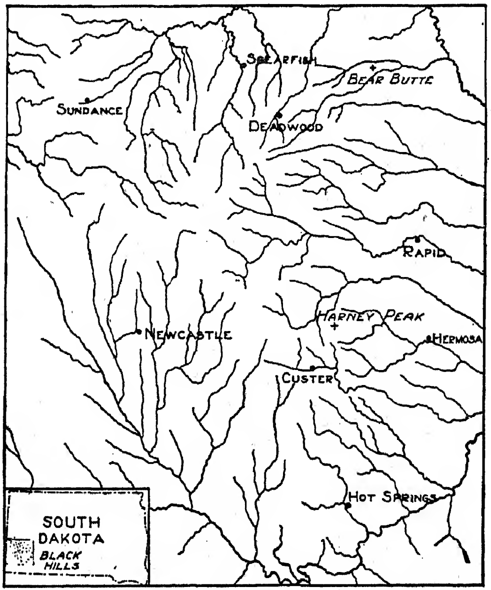
The development of flats commonly accompanies the widening of a valley, especially where the stream has cut its channel down to a low gradient (Fig. 94). Such flats are temporary base levels established by the local conditions that govern the stream's gradient. Later, when conditions of velocity, volume, and load are altered, the stream may cut deeper into the flats and leave them as terraces along the valley. The Mississippi River has developed flats along its upper course where the floor of the valley is more than 600 feet above sea level (Fig. 95). A short distance south of St. Paul, it has such a flat over a mile wide that is cut about 250 feet below the general level of the region. In the neighborhood of St. Louis the flats are about 400 feet above sea level and about 150 feet below the regional upland. Still further downstream in Tennessee and Arkansas the flats are 35 miles wide and lie 220 feet above the sea.

The width limit of a valley depends on the distance between neighboring streams. Parallel valleys may be widened until the divide between them becomes a sharp, narrow ridge that is gradually lowered as the region approaches base level.

Trend of the Stream's Course.—There are various causes which lead to the selection of the course followed by a stream in its journey toward the sea (Figs. 96 to 99). A valley may owe its position and its trend to one or more of the following factors: (1) the original slope and natural



96.—A dendritic drainage pattern characteristic of an area underlain by horizontal strata or massive crystalline rock of uniform hardness.



97.—A radial stream pattern developed on the surface of the dome-like uplift of the Black Hills in South Dakota.

irregularities of the surface, (2) differential erosion, (3) jointing, (4) faulting, and (5) folding.

1. A river whose course has been determined by the original slope and the irregularities of that slope is a *consequent stream* (Fig. 100). Such streams are characteristic of coastal plain areas where the surface is comparatively uniform and regular, with a gentle slope to the sea. Many of the streams along the Atlantic and Gulf coasts of the United States are of this type.

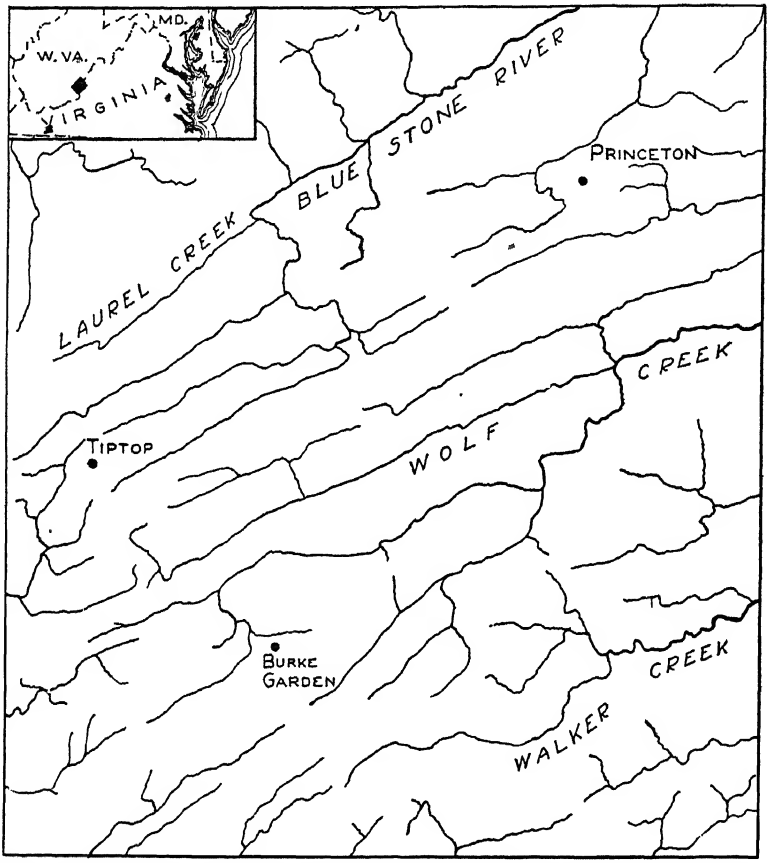


FIG. 98.—An adjusted stream pattern characteristic of the mature stage of an erosion cycle in a region of folded or tilted strata of different degrees of resistance to erosion.

2. As erosion proceeds, new channels are developed, independent of the original topography, and their courses are directed by differences in the structure and character of the bedrock formations. The streams tend to follow the softer beds. Such variations lead to differential erosion so that eventually a stream may undergo marked changes in position and direction and alter its original consequent course. Rivers formed in this way are called *subsequent streams* (Fig. 101) because they

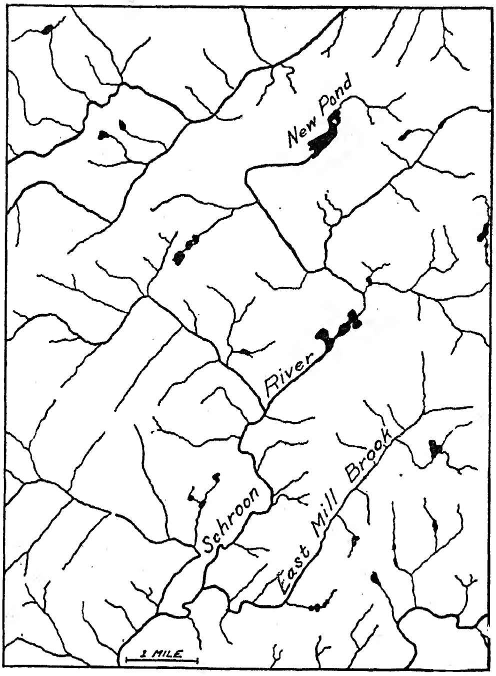


FIG. 99.—A rectangular stream pattern near Elizabethtown, New York. The pattern is characterized by right-angled bends in both the main stream and its tributaries. It is controlled by right-angled jointing or faulting of the rocks.

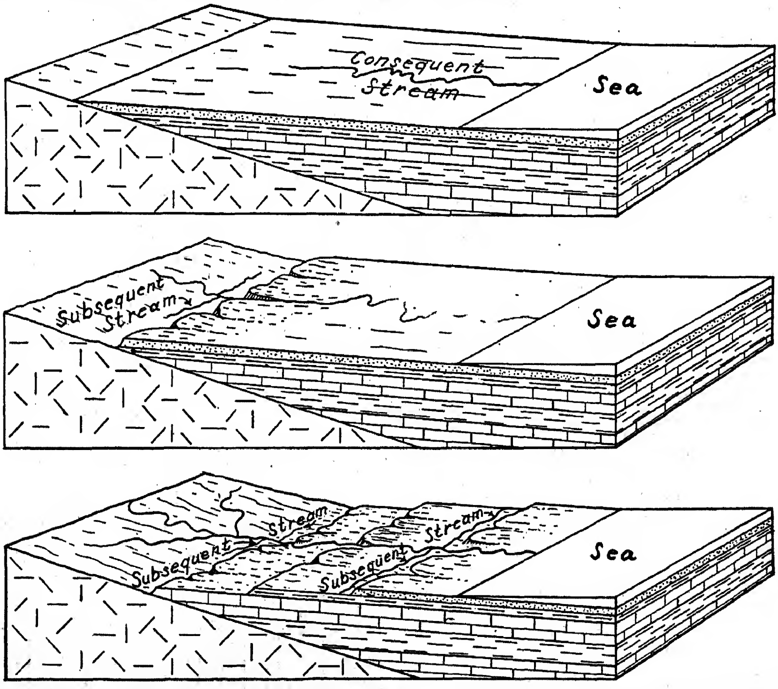


FIG. 100.—A series of diagrams showing the development of subsequent streams.

have been developed by subsequent erosion determined by structure. Topographic features produced by differential erosion are described in a following section.

3. The position of a valley often is controlled by the direction of the joints or fissures in the bedrock of the area which it drains (Fig. 99).

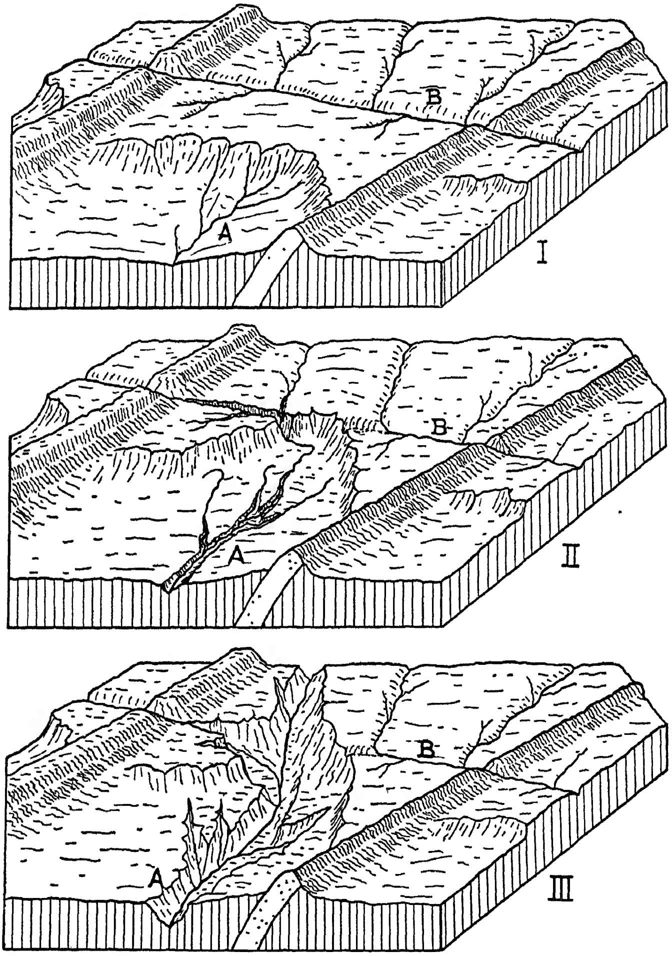


Fig. 101.—Diagram illustrating stream piracy. *I*, the tributaries at *A* are advancing by headward erosion toward the valley of the stream *B*. *II*, the stream *B* has been beheaded or captured and its headwaters are diverted to the pirate stream *A*. *III*, the valley of *A* is extended and deepened. (*Based on a drawing by Davis.*)

This is true especially of smaller tributary streams, as is characteristically shown in Connecticut and also in Ontario where large areas of strongly jointed metamorphic rocks are exposed at the surface.

4. In regions where faulting has taken place on a large scale, many valleys follow the fault planes for great distances. At some places long, narrow blocks of the earth's crust have been depressed to form

valley-like basins, which later become stream channels. The Dead Sea basin and the Jordan valley are typical examples. In California, Owen's valley has a similar history; and in Germany a large portion of the valley of the Rhine is a structural trough flanked by the Vosges and by the Black Forest mountains in which many of the steep slopes facing the valley are escarpments produced by the displacement of the rocks along fractures.

5. Long, parallel mountain folds influence the trends of valleys by governing the directions of the consequent streams which follow the troughs of the folds. Small streams flow from the crests of the folds into the troughs where they unite to form the larger ones. At such places the walls of the valley are the limbs of the folds; and since the courses and profiles of the valleys are determined by the structure of the rocks through which they pass, they are *structural valleys.* Certain tributaries of the Columbia River in Central Washington follow such valleys.

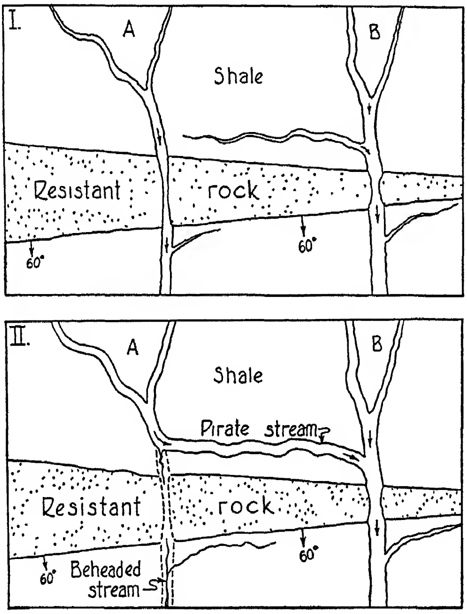


Fig. 102.—*I*, an area of highly tilted rocks with differences in resistance to erosion. Two streams cross the beds. Stream *B* is flowing over the resistant rock where it is thinner than at stream *A* and therefore it deepens its valley more rapidly than *A*. *II*, through headward erosion the tributary of stream *B* has captured the headwaters of stream *A*.

Stream Piracy.—In the process of valley development each stream continues to extend or to modify its drainage basin until all of its divides become stationary. During this process it frequently happens that one stream finds conditions for growth and extension more favorable than another, on the opposite side of the divide, and, by the extension of tributaries, cuts back until it steals some of the headwaters of the less favorably situated stream and diverts them to its own channel. Such invasion is *stream piracy*, and the stream whose territory has been invaded is said to have been *beheaded* by the pirate stream (Figs. 101, 102). The conditions which may give a stream an advantage over an adjoining one are (1) a greater volume of water, (2) softer rocks in which to excavate its channel, and (3) higher gradient due to a shorter course to the sea. If the amount of precipitation is greater on one side of a mountain range than on the opposite slope, the streams receiving the greater volume of rainfall will have a higher velocity and therefore will erode more rapidly and will be able to extend their headwater tributaries farther than streams on the opposite side of the divide.

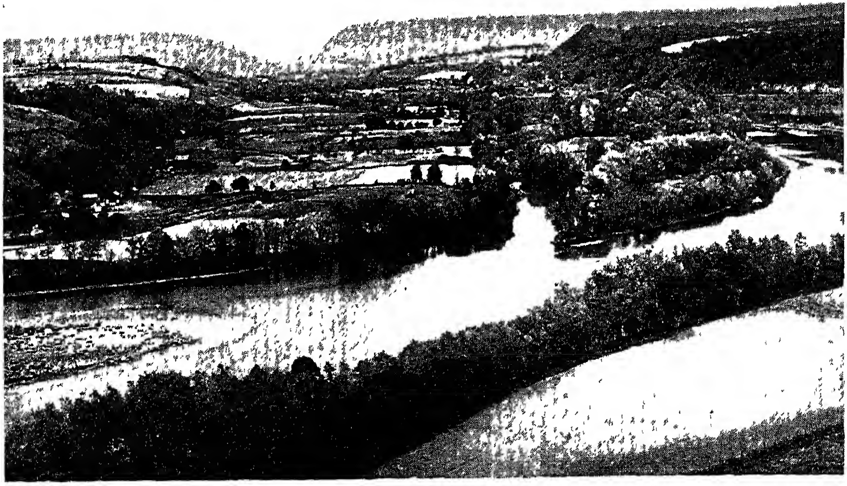


FIG. 103.—The flood plain, terraces, and water gap of the Delaware River. (*Photograph by Stose, U. S. Geol. Survey.*)

If streams drain regions with tilted strata in which different types of sedimentary rocks alternate as a result of tilting, the larger streams tend to follow the outcrops of the less resistant beds, and their smaller tributaries join them nearly at right angles. Where such conditions prevail, the larger streams flowing in the softer rocks often behead the streams that cut across the hard strata. During high water the higher stream overflows into the valley of the deeper stream and establishes a channel which later becomes permanent.

Where a stream flows across tilted strata, a *narrows*, or a *water gap* (Fig. 103), is developed where the valley crosses the harder beds. If a stream is diverted from the water gap by piracy, the narrow portion of the beheaded valley is called a *wind gap*. Such gaps are common in the

Appalachian region, and at many places they served as passes through the mountains for the early pioneers traveling by wagon to settle in Kentucky and Tennessee. It is estimated that 300,000 people passed through the Cumberland Gap in their migration westward during the last quarter of the eighteenth century. Some of the gaps of the Blue Ridge Mountains became strategic points during the campaigns of the Civil War.

STAGES IN VALLEY DEVELOPMENT

Cycle of Erosion.—A newly uplifted land area marks the starting point of the history of a river-drainage system. It is evident that the topography produced by stream erosion, as such an area is drained will change as erosion continues. Changes are determined in part by the altitude of the uplifted area and the character and structural relations of its subsurface rock formations. The first effect of erosion is to roughen the surface of the area by excavating gullies and valleys and by leaving ridges and hills as divides (Fig. 104). Eventually, the divides are lowered, and the result is a plain or comparatively level surface. Hence, erosion tends eventually to produce a plain. The time involved in the development of the various stages of topographic expression through which a region passes as it is transformed into a featureless plain comprises a *cycle of erosion* (Figs. 105, 106). For purposes of comparison and study of land surfaces, this cycle is divided into three stages—youthful, mature, and old. The topography of a valley is youthful when most of the work of the stream is not yet done; it is mature when the erosive agents have accomplished so much work that most of the area consists of slopes; and it is old when erosion has nearly finished its work. The use of these terms is intended to convey the idea of stages rather than periods of time expressed in years; for one valley may reach maturity in a much shorter period than an adjoining one, especially if it is in softer rocks and the latter in a region of more resistant rocks. A distinction should be made also between the age of the stream in the cycle of erosion and the stage in the erosion cycle to which the area it drains has progressed, for certain streams with youthful characteristics may traverse regions that are typically mature.

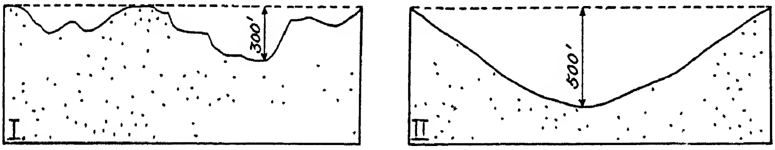


Fig. 104.—Cross sections showing: *I*, rough surface with little relief; *II*, smoother surface with higher relief.

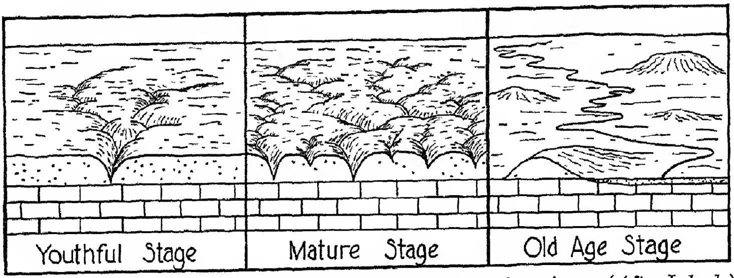


FIG. 105.—Illustrations showing the stages in a cycle of erosion. (*After Lobeck.*)

Summary of Changes during a Cycle of Stream Development

Characteristics	Youthful stage	Mature stage	Old-age stage
Trends of channels	Straight	Meanders common	Meanders numerous
Gradient and velocity	High	Moderate	Low
Waterfalls and rapids	Many	Few	None
Nature of erosion....	Downcutting predominates	Lateral planation prominent	Lateral planation predominant
Width of valleys....	Narrow, V-shaped	Broad and well defined	Very broad with low boundaries
Depth of valleys....	Moderately deep	Deepest	Shallow
Number of tributaries	Few, small	Maximum number	Few, large
Nature of divides....	High and wide	High and narrow	Low and narrow
Relief..............	Maximum for entire drainage system	Maximum for region of headwaters	Low
Number of lakes.....	Many on uplands	Very few	Many on lowlands
Adjustment to structure..............	Not adjusted	Well adjusted	Roughly adjusted
Material transported	Coarse and fine	Sands and silts prominent	Sands and silts predominant
Deposition by streams	Minimum deposition	Deposition at insides of curves	Deposition in channels and on levees
General drainage....	Poorly developed	Well drained, most efficient	Drainage sluggish

Youthful Streams and Valleys.—Most youthful streams are rapid streams that flow in V-shaped canyons or gorge-like valleys with steep sides. The slopes are steep because sufficient time has not yet elapsed for the valleys to be widened. If the area is a recent uplift, the streams are not numerous, and they have few tributaries. Since they have not had time to erode extensively, they still may have rapids and waterfalls along their courses. The divides are wide and are poorly drained, as is shown by the presence of upland lakes and swamps. This condition is illustrated in the valley of the Red River of the North, or in portions of the coastal plain areas of southeastern United States. The gorge of the

Niagara River, the canyon of the Yellowstone, and the valley of the Rhine are all youthful valleys. The Grand Canyon of the Colorado, which also is a young valley, is the grandest of all examples.

Maturity.—As erosion continues, the topography changes until the features characteristic of youth are chiseled into different forms, and the sharp, straight lines of the landscape give place to valleys with flaring sides and gently rounded upper slopes. By the gnawing back of the headward tributaries, the divides are narrowed, and the region becomes thoroughly dissected by a complex network of valleys resulting in very rugged topography. The number of tributaries is to some extent determined by the amount of rainfall. In southeastern United States with a heavy annual precipitation, tributaries and streams are more numerous than in the semiarid or arid plains of the southwest. As the tributaries are deepened, the lakes and swamps are drained or filled, and the escarpments that produced waterfalls and rapids are lowered, so that the rivers attain gradients that are near the lowest slopes over which their loads of sediment can be transported. Toward the close of this stage the lower courses of the streams become graded, and then they swing from side to side, thus widening their valleys and developing flood plains along the valleys.

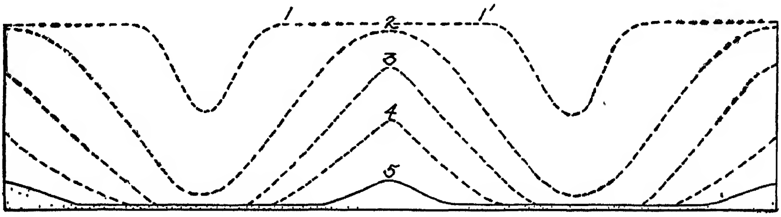


Fig. 106.—Diagram showing change of profile of a divide during an erosion cycle. As the youthful V-shaped valleys are widened, the intervening upland 1-1′ is narrowed. The mature profile 3 shows the divide narrowed to a ridge. As erosion continues the ridge is lowered to profile 5 and wide valley flats are developed. Such a profile is typical of old age.

From the standpoint of human activities, rugged mature topography offers many obstacles; and it is the least desirable for many enterprises. Because of the network of deep valleys, roads cannot follow the straight survey lines but follow the crests of the main divides or the winding valleys of the major streams. Railroad grades cross areas of mature topography by using high trestles over the valleys and by penetrating the hills with long tunnels. Typical mature topography now exists in the region of the Allegheny and Cumberland plateaus to the west of the Appalachian Mountains.

Old Age.—By continued erosion the rugged relief of the mature stage of topography gradually is reduced, and the deep channels are transformed into broad valleys with gentle slopes and low divides. The gradient is lowered until the streams lose their vigor and deposit rather than erode. The valleys become shallower, owing to deposition, and the sluggish streams swing from side to side in long loop-like meanders over the deposits of their own flood plains. A land area thus worn down to nearly a plain with the gentle slopes of old-age topography is a *peneplane* (Fig. 107). Frequently, isolated hills or mountains of more resistant rocks rise to greater heights above the general level of the peneplane. These are called *monadnocks* after the type example of Mount Monadnock in New Hampshire (Figs. 108, 109).

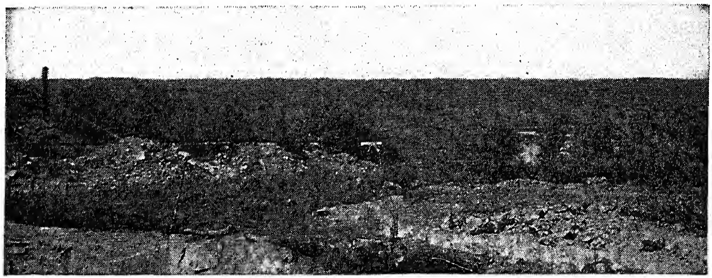


Fig. 107.—Photograph showing a profile of the Laurentian peneplane of Canada. The sky line is flat. (*Photograph by Hyde.*)

Relation of a Peneplane to Base Level.—Theoretically, the ultimate base level of all stream valleys is sea level. Since the flow of a stream is influenced by gravity, however, it is evident that erosion can proceed only as long as a stream retains sufficient gradient to transport its load. As a land surface approaches a peneplane, it approaches also base level. The base level of stream erosion is the lowest possible level to which an area may be eroded by running water. In many regions a peneplaned condition is produced while the region still is several hundred feet above sea level.

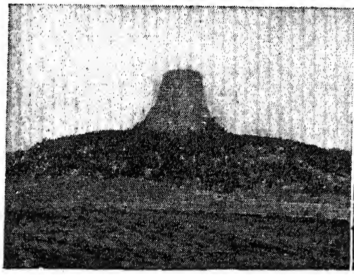


Fig. 108.—A monadnock. Devil's Tower, Wyoming. (*Photograph by Stephens.*)

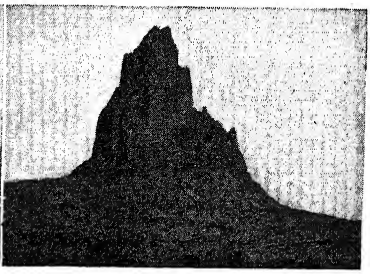


Fig. 109.—A monadnock near Flagstaff, Arizona. (*Photograph by Colton.*)

Disturbed Erosion Cycles.—As the drainage basin of a river system passes through the successive stages of a normal cycle of erosion, it may be interrupted at any stage by various geological processes. Among them the following are perhaps the most important: (1) glaciation, (2) volcanic action, and (3) diastrophism.

1. A glacier may fill a stream valley with ice or cover its basin with a snow field that protects the surface from weathering and stream erosion. During the great ice age when most of northeastern North America was subjected to continental glaciation, many areas within that region were in the mature stage of topography. As the glacier receded, however, enormous amounts of glacial débris were deposited irregularly, forming hills, ridges, knobs, and depressions over the preglacial topography, with the result that youthful features again were developed and superimposed on large areas which had been eroded far beyond the youthful stage. This is true especially of the region to the south and east of the Great Lakes in the United States.

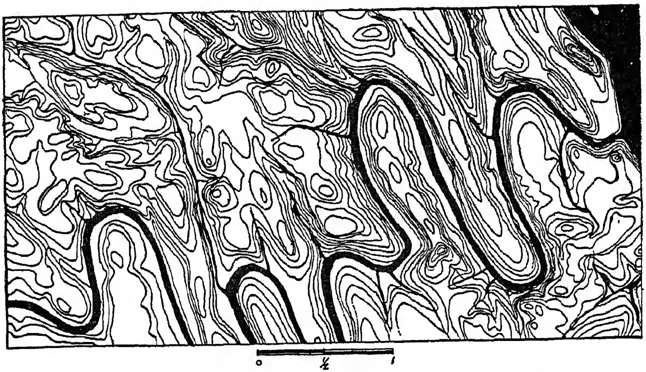


FIG. 110.—A contour map of a rejuvenated stream with intrenched meanders. The meanders were inherited from the old stream which preceded rejuvenation. (*Harrisburg quadrangle, Pennsylvania, U. S. Geol. Survey.*)

2. Lava may flow down the slopes of a valley and completely fill its stream channel, or the volume of lava may be so great that the entire area is deeply buried, and the drainage system profoundly modified. In such areas the erosion cycle is interrupted and a new cycle is initiated on the surface of the lava field. The Columbia plateau, covering an area of over 200,000 square miles in Washington and Oregon, is a lava field where the streams and valleys of the old erosion surface were at many places almost completely buried.

3. Diastrophism may bring about interruptions that are varied both in character and in the results produced. They may modify the cycle of erosion only locally and temporarily, or they may cause the streams to start their work anew. Changes of level may be produced either by elevation or by depression.

Rejuvenation.—If a peneplaned area is elevated, the gradients of the streams are increased, and they set to work cutting gorges and canyons in the bottoms of their old valleys, with the result that the region again takes on the characteristics of youth. Such a region is said to be rejuvenated; the infusion of new vigor through an increased gradient quickens the velocity of the streams, and they become rejuvenated streams.

FIG. 111.—Intrenched meanders in San Juan Canyon, Utah. The meandering course was developed during a former erosion cycle, when the river was flowing over a graded plain. The meanders became intrenched when the stream was rejuvenated by the uplift of the area. (*Photograph by H. H. Vinson.*)

After such an uplift, a river sinks its valley within the new upland; and if the old streams had meandering courses before rejuvenation, the winding channel is deepened, and the old meanders become intrenched[1] (Figs. 110, 111). Eventually, during renewed maturity, the wide lobes of the individual meanders become separated by mountains formed by the dissection of a slowly elevated plain. If, during the elevation, a stream is able to hold its course notwithstanding change of

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[1] Four different methods are used to represent relief on a flat sheet of paper or on a map. These are: by color, by shading, by hachures, and by contours. The contour method is the commonest and most nearly accurate. A contour line on a map represents an imaginary horizontal line that passes through all points on the surface of the area mapped that have the same elevation above a given datum—usually sea level. The lines are drawn to indicate a certain regular difference in elevation, and this difference is called the contour interval. The accompanying illustration (Fig. 112) represents a contoured conical hill. The contour interval is 10 feet. Contours bend, or "loop," upstream in valleys. They are closely spaced on steep slopes and distantly spaced on gentle slopes. They encircle conical hills and are paired on opposite sides of parallel ridges. The contours near the crests of hills are relatively short, closed curves.

level in the earth's crust, the stream is *antecedent*—that is, it antedates the diastrophic events that influence the present topography.

Antecedent streams with intrenched meanders are common in the Appalachian region. The Susquehanna River of southern New York and northern Pennsylvania and the New River of the Cumberland Plateau are examples. The Yakima River of central Washington and the San Juan River of southeastern Utah are both deeply intrenched. Some of the natural bridges of Utah owe their origin to the perforation of necks between intrenched meanders. The famous Rainbow Bridge, carved out of sandstone, has a height of 305 feet and a span 270 feet long across the neck between old meanders (Fig. 140).

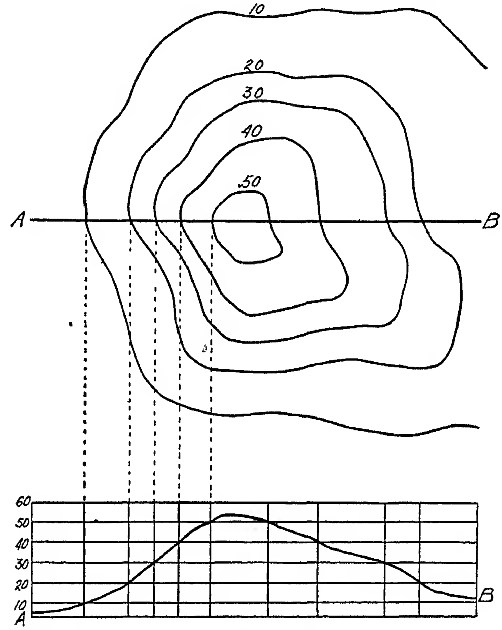

FIG. 112.—Diagram showing a contoured hill. The contour interval is 10 feet. The lower diagram is a profile of the hill along the line *A-B*.

Uniform slow emergence, or elevation, affecting extensive areas may revive the streams without greatly altering the major features of the topography. In this way an old peneplane may be lifted up to an altitude of several thousand feet above sea level without warping the old erosion surface.

Faulting may produce tilting; and where the movement is upward, the streams are revived, and the cycle of erosion is interrupted. If the movement is downward, the gradient is decreased, and the velocity of the stream is checked. Notable depression or elevation along a fault line across a valley may so impede the flowing streams that the waters become *ponded*, and the rate of erosion is changed.

Depression of the land hastens the development of old age by bringing the depressed area nearer base level and by decreasing the amount of material the streams must remove before a featureless plain is developed. When such subsidence takes place along the coast line, the sea occupies the lower ends of the valleys and converts them into bays and estuaries. The valleys then are *drowned.* Thus the Hudson River is drowned as far north as Albany, and the St. Lawrence River appears as an arm of the sea as far as Montreal. Narragansett, Delaware, Chesapeake, and other bays along the coast of the United States are drowned valleys resulting from a sinking of the seaboard areas. Before submergence took place, many of the streams which now flow into Chesapeake Bay discharged their waters into the lower course of the Susquehanna River, to which they were tributaries (Fig. 113). Such tributaries, isolated from their trunk streams by drowning, are called *dismembered* streams. With subsequent emergence they would extend their courses and again become parts of the major stream system.

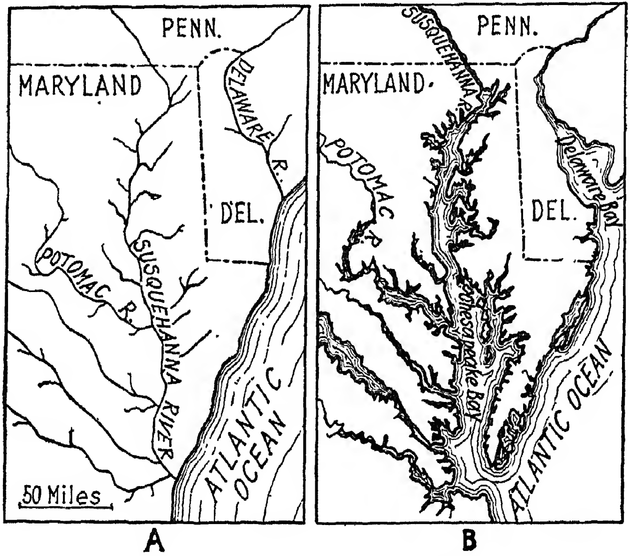

FIG. 113.—Drowned and dismembered rivers. Delaware and Chesapeake bays (*B*) were formed by sinking of the coastal plain, which permitted the Atlantic Ocean to drown the lower parts of the valleys. The tributary rivers (*A*) which formerly emptied into the main streams are *dismembered* streams. *A*, probable conditions before subsidence; *B*, present conditions. (*Based on a drawing by Salisbury.*)

DIFFERENTIAL EROSION AND STREAM ADJUSTMENT

Stream Erosion in Flat-lying Strata.—Where the strata are horizontal or but slightly inclined, the valley form is determined by the nature of the rock formations. If excavated in strata of uniform resistance, the valley slopes show few irregularities, and the angle of the slope is determined by the ratio of lateral erosion to the deepening of the channel. As erosion continues, successive layers are exposed, and in an area of maturely dissected topography the outcropping beds swing out around the spurs between tributaries and up into the tributary valleys. If followed upstream on one side, a given stratum at the level of the stream crosses the stream and turns back on the other side (see Fig. 114).

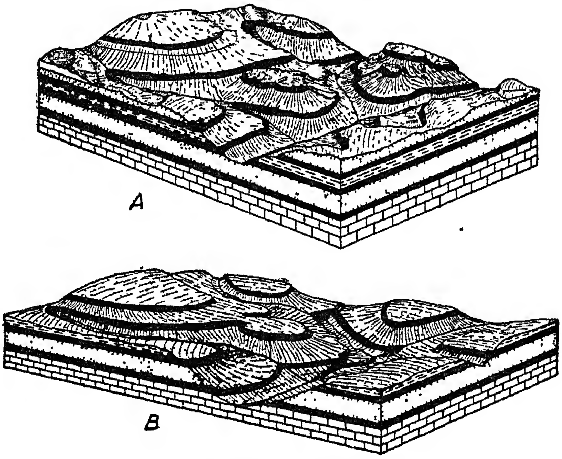

Fig. 114.—Diagrams showing the erosion of flat-lying sedimentary strata. In *A*, the softer strata are worn away more rapidly than the harder (shown black), which remain as steep cliffs. In *B*, erosion has proceeded further, the deep gullies and sharp, angular profiles have been smoothed off, and the valley slopes made more gentle. (*From a model by Ward's Natural Science Establishment.*)

Rock Terraces.—By differential weathering, the resistant layers are etched into relief, and the slopes of the valley become terraced (Fig. 115). Such structures are *rock terraces*, since they are cut in solid rocks rather than in alluvium.

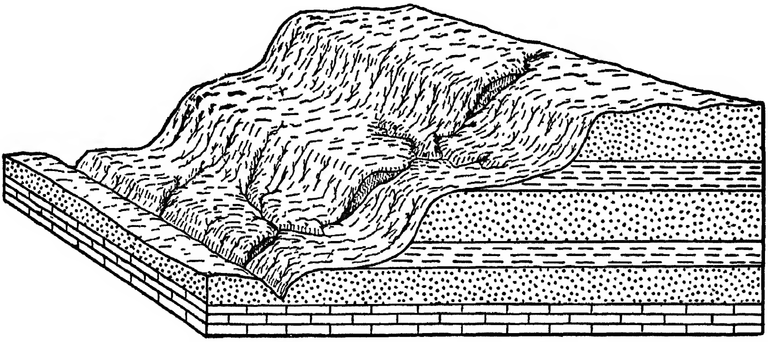

Fig. 115.—Diagram showing the development of rock terraces. The harder sandstone layers through which the main stream has sunk its valley remain as rock terraces along the side of the valley. Steep gullies form where a tributary valley crosses the terraces.

The downward slope of a rock terrace is the exposed edge of a hard stratum and usually is steep and cliff-like, whereas the slope rising above the terrace is formed by the eroded edge of the softer stratum above and normally is a gentle slope covered with weathered rock waste. If the resistant beds are thick, extensive escarpments may develop, as is shown on a magnificent scale in the Grand Canyon of the Colorado River.

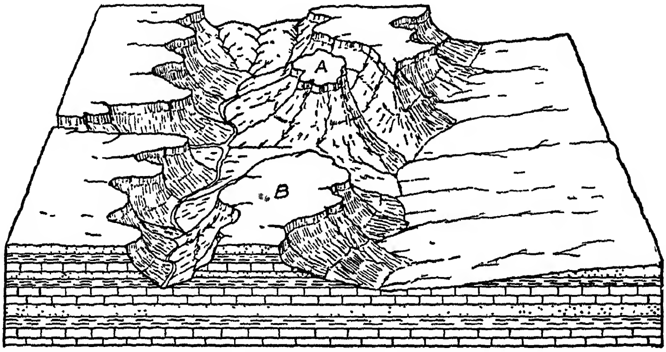

Fig. 116.—Diagram showing the development of mesas and buttes by stream erosion. *A*, butte; *B*, mesa. (*Based on a drawing by Brown and Debenham.*)

As the topography becomes more mature, the terraces cut back farther from the channel, and broad, flat areas many miles in width and parallel to the stream may be developed on both sides of its valley.

Mesas and Buttes.—In regions of horizontal sedimentary strata or where sheets of lava cover soft clays or partially indurated sediments,

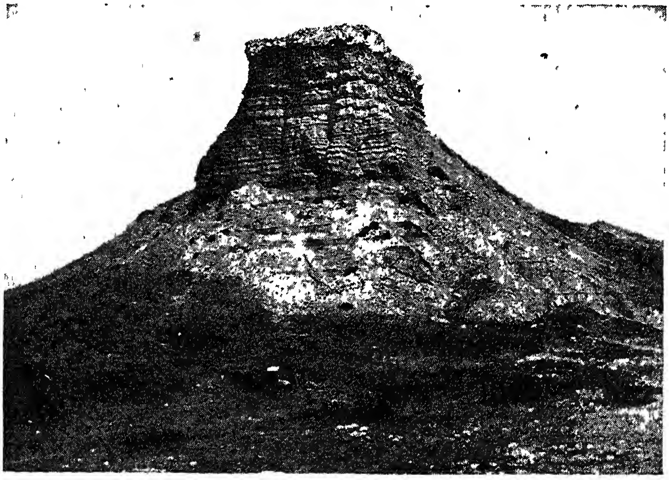

Fig. 117.—Red Butte, Wyoming. (*Courtesy U. S. Geol. Survey.*)

lat-topped areas are isolated by the headward cutting of tributary streams as the region passes from youthful to mature topography. Such plateau-like areas are *mesas,* from a Spanish word meaning table. The evel top of a mesa consists of a resistant horizontal bed that tends to protect the less resistant strata below it. Where erosion reduces a mesa to a flat-topped hill, it becomes a *butte* (Figs. 116, 117). Many of the striking mesas and buttes in the semiarid plains of New Mexico and Arizona are remnants of former plateaus dissected by erosion.

Bad Lands.—Under special conditions the differential erosion of flat-lying beds produces peculiar and striking types of topography (Fig. 118). Among these are the *bad lands* which are well developed in the northwestern Great Plains region of United States and Canada, especially the Dakotas, Wyoming, and Montana. Bad lands have flat-topped, or butte-like, hills with rugged, barren slopes. The ruggedness is due chiefly to numerous gullies cut in the slopes of slightly consolidated sediments, in which certain layers are sufficiently resistant to erosion temporarily to arrest denudation. A semiarid climate is favorable to the development of such features, for under such conditions the rainfall is concentrated in a few heavy showers, and thus there is the maximum run-off and hence the maximum of erosion with the minimum rainfall. The semiarid climate is largely responsible for the lack of vegetation on the slopes which thus expose the loosely consolidated material to the torrents of water.

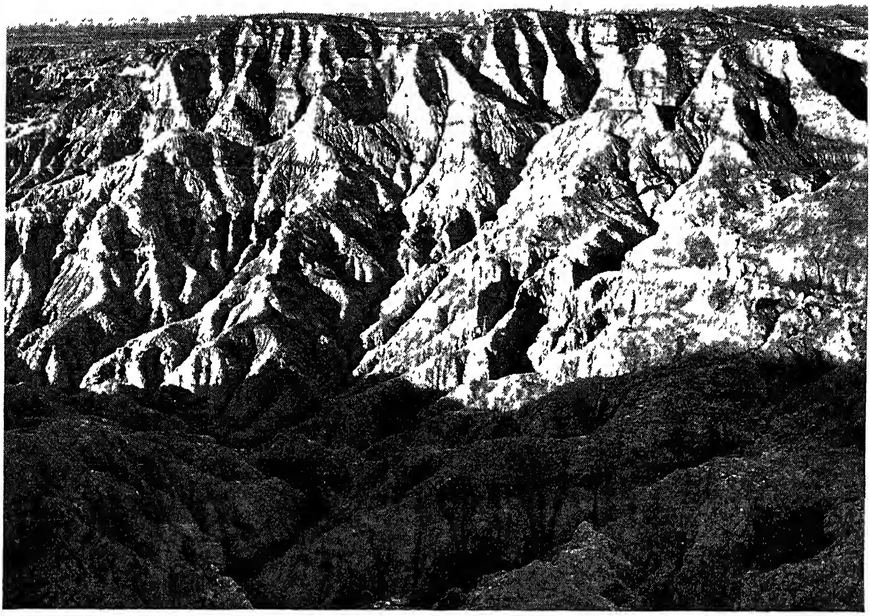

Fig. 118.—Bad-lands topography near Scotts Bluff, Nebraska. (*Photograph by Darton, U. S. Geol. Survey.*)

Erosion of Tilted Strata.—Where a region underlain by tilted strata (Fig. 119) is eroded toward maturity, the inclined beds are exposed, and some of the alternating layers are more easily eroded by the streams than others, with the result that they tend to flow, as far as possible, on the less resistant beds and to cut across the resistant ones, thus flowing as short a distance as possible on resistant beds. The less resistant strata, therefore, become the sites of valleys, and the resistant strata stand up as ridges or mountains. Such changes in the course of a stream by means of which it develops a definite and stable relation to the subsurface rock structures are called "structural adjustment."

FIG. 119.—Erosion of inclined strata exposing dip slopes. Creation Rock, Red Rocks Park, near Denver, Colorado. (*Courtesy Denver Tourist Bureau.*)

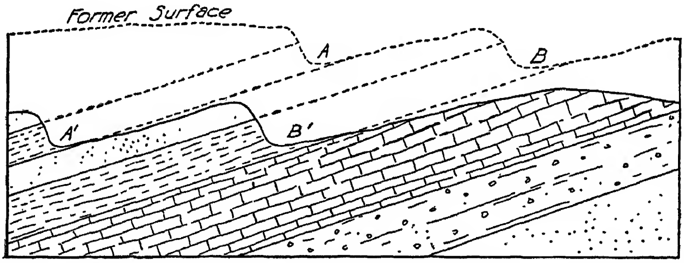

FIG. 120.—Diagram showing monoclinal shifting of valleys. As the earth's surface was lowered from the upper profile shown by the dotted line, to the profile shown by a continuous line, valleys *A* and *B* moved down the dip of the strata to *A'* and *B'*.

The Hoosic valley of Massachusetts, the Shenandoah valley of Virginia, and the Lehigh valley of Pennsylvania are all examples of valleys that are eroded in relatively soft rocks.

The adjusted streams, flowing parallel to the strike, frequently cut vertically through the weaker strata until they encounter a hard or resistant bed and then shift down the dip, developing a valley with a dip slope on one side and a steep escarpment on the other (Fig. 120). Such lateral shifting down the dip is *monoclinal shifting,* and the valley is asymmetrical because of the difference in profile of the two slopes. Where parallel streams flow at right angles to the strike of slightly inclined beds, the strata crop out in parallel belts with sharp bends.

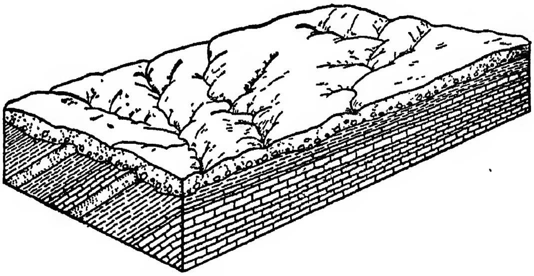

FIG. 121.—Diagram showing valley development in a flat-lying stratum that rests on the eroded edges of folded strata. Compare with Fig. 122.

Where tilted beds are peneplaned and covered by later flat-lying beds, the drainage system will be independent of the structure of the tilted beds. Where such a system by continued erosion is let down on the surface of the tilted beds, the streams may cut across them. Such streams are *superimposed.* (Figs. 121, 122, 123.)

Cuestas.—Where warping has tilted the rock strata to low angles, as along the Atlantic coastal plain where the beds dip seaward from 5 to 12 degrees, erosion develops ridges on the more resistant layers, and from the crest of such a ridge there is a steep slope facing landward and a gentle descent toward the coast. Such a ridge with steep erosional escarpments on one side and dip slopes on the other is a *cuesta.* Where there are alternating beds of hard and soft formations, there will result alternating zones of lowlands and highlands in roughly parallel bands. An example is the Black Prairie of Alabama, bordering the Appalachian Mountains, consisting of a belt of lowlands formed over easily eroded

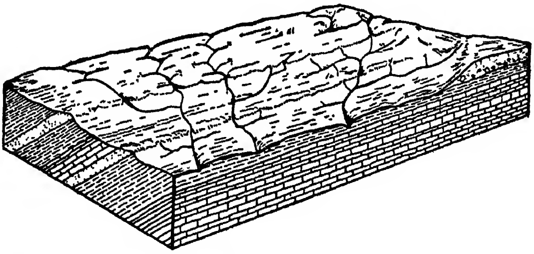

FIG. 122.—Diagram of the region shown in Fig. 121 after erosion has removed the horizontal bed. The streams are not in structural adjustment because the old drainage is superimposed upon the tilted strata.

sediments. At the margin of the prairie is a cuesta which faces the lowland and rises abruptly about 200 feet from the surface of the lowland.

Hogbacks.—Where strata are inclined at high angles (Fig. 124), the ridges produced by differential erosion have slopes that are nearly equal. The side corresponding to the cuesta escarpment develops a more gentle slope, and the dip slope of the opposite side is steep. A ridge with such a profile is a *hogback*. They are commonly developed on the flanks of folded mountains such as those of the Rocky Mountain system, where they are conspicuous especially along the east margin of the Front Range in Colorado and Wyoming. Eastward toward the plains there are all gradations from hogbacks with steep slopes formed in highly tilted beds, through cuestas formed where the inclination is low, to mesas and buttes in the horizontal strata of the plains; and all are simply remnants of erosion that assume different profiles as a result of the different attitudes of the rock formations.

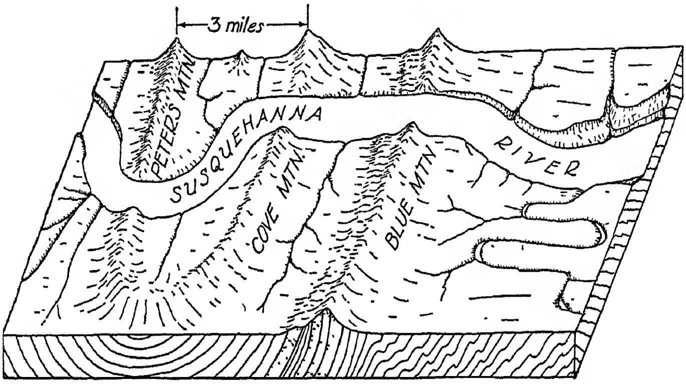

Fig. 123.—A superimposed stream flowing over truncated folds. The Susquehanna River near Harrisburg, Pennsylvania. (*After D. W. Johnson.*)

Fig. 124.—Diagram showing the outlines of various physiographic forms produced by differential erosion of tilted and of horizontal beds.

Erosion of Folds.—On symmetrical folds, such as a series of parallel anticlines and synclines, the initial consequent streams flow in small gullies on either side of the crests of the anticlines and drain away from their axes. These small streams discharge into the synclines, where the major streams flow along the axes of the synclinal troughs. As tributaries cut back into the flanks of the anticlines, they develop gorges that soon become sufficiently deep to develop lateral gullies of their own, and working in rocks that were fractured during folding these tributary gullies rapidly cut deep into the axial portion of an anticline where, in many of them, numerous fractures and joints hasten the process of denudation. In time, the divides between the tributaries that flow on the crests of the folds are narrowed and lowered, and lateral erosion widens their valleys until the valleys are pushed down the flanks of the anticlines. Stream conquest or piracy between adjoining folds follows, and eventually the streams along the crests of the anticlines become the master streams, and the major valleys become anticlinal valleys (Fig. 125). The synclinal divides stand high in relief and form broad ridges and, at places mountains. Lookout Mountain is an example of a synclinal divide which terminates at the north in a steep escarpment 1,500 feet high at Chattanooga, Tennessee, and extends southwest over 50 miles into Alabama. A deep anticlinal valley from 4 to 5 miles wide flanks Lookout Mountain on the west, and still farther westward beyond this valley is another broad synclinal plateau.

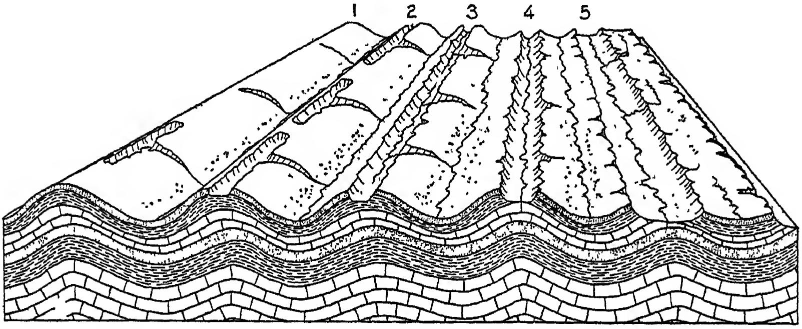

Fig. 125.—Diagram of a series of eroded folds showing the development of anticlinal valleys and synclinal divides. The maximum fracturing is along anticlines. Fold 1 shows three tributaries that flow into the synclinal valley between folds 1 and 2. In fold 2 the tributaries have developed valleys along the axis of the anticline. In fold 3 the tributary in the foreground has captured the headwaters of the second tributary, and in fold 4 all of the tributaries have been captured by one downstream beyond the diagram. In fold 5 the stream has cut through the resistant sandstone on the crest of the anticline and therefore it can erode downward more rapidly than the stream in the syncline between 4 and 5. Eventually it diverts the water of the stream in the syncline as shown by the tributaries to the right of 5. Thus the syncline becomes a divide. (*After Lobeck.*)

Curved Valleys and Zigzag Ridges.—Where symmetrical folds pass through a cycle of erosion, the resistant layers form parallel ridges that are paired on the two limbs of a fold. If, however, the axis of a fold is not horizontal but is tilted so that it plunges into the earth, the erosional ridges present a different topographic pattern. If the folds plunge steeply some of the pairs of ridges end after the layers producing them plunge below the regional base level. When the plunge is less steep, they converge in the direction of plunge and eventually join to form a continuous ridge with a sharp, elbow-like flexure and with a valley shaped like one end of a canoe within the loop of the flexure. If several resistant layers are present, each will form an encircling ridge with curved valleys separating them. A series of alternating plunging anticlines and synclines, therefore, produce zigzag ridges (Fig. 126), such as characteristically are developed on the Appalachian peneplane in Pennsylvania, New York, and Virginia.

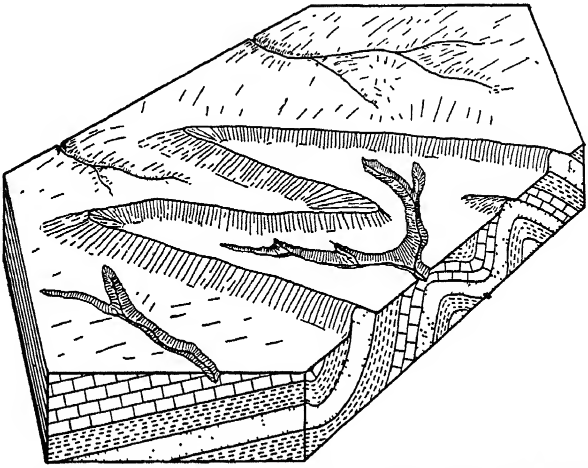

FIG. 126.—Zigzag ridge produced by the erosion of plunging folds.

Erosion of Domes and Basins.—On a newly uplifted dome-like structure the initial streams form a series of radial valleys extending in all directions from the crest of the dome. These *consequent* streams may unite to form one or more trunk streams at the lower margin or base of the structure. As erosion cuts through the strata on the summit of the uplift, the formations begin to crop out in a series of narrow belts around the crest of the dome (Fig. 127). If beds of varying hardness are present in a series of sedimentary strata, the harder sandstones and limestones form ridges, and streams adjust themselves to the softer, shaley layers. In this way the radial valleys are converted into concentric, or *ring*, valleys, and the ridges are transformed into hogbacks which stand concentrically around the center of the uplift. Because of the steep escarpments facing the valleys, some of the more conspicuous sandstone ridges rim the valleys. Examples are found in Montana and South Dakota.

As structural basins pass through a cycle of erosion, they tend to show similar concentric features but with the escarpments facing outward. The structural relations are similar to those observed in a low nest of shallow plates, in which the largest is placed at the bottom and the smallest at the top. In such a nest the outer edges of the plates correspond to the encircling ridges of harder rock that rim the basin as the region is base-leveled. The lower peninsula of Michigan and the Paris basin in France show such structural relations.

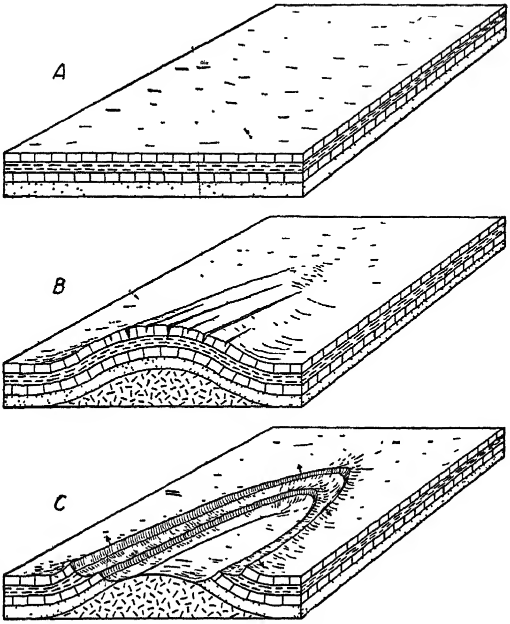

Fig. 127.—A series of diagrams showing the formation and dissection of a dome-like uplift. *A*, horizontal strata; *B*, strata arched upward and fractured by an intrusive mass of igneous rock; *C*, dome maturely dissected, exposing the igneous core and concentric hogback ridges sculptured from the more resistant strata.

FEATURES DEVELOPED UNDER SPECIAL CONDITIONS

Waterfalls and Rapids.—The geologic significance of waterfalls may be slight, but from a scenic point of view they form a fascinating part of a river system. There is no sharp distinction between a rapid and a fall. Steep rapids are commonly called falls, and, when small, both are referred to as cascades, and where an enormous volume of water falls over a precipice, the term cataract is used.

Falls and rapids occur at many places, and are formed under various conditions. Wherever the bedrock is made up of layers that have different degrees of resistance to erosion, the resistant layers hold back the streams in their trenching processes. Since the less resistant beds farther downstream are still being cut away at the old rate, there is a lack of adjustment between that part of the stream above and that which lies below the outcrop of the resistant layers. There are thus developed a series of rapids which get steeper and steeper with continued erosion and finally become a waterfall. Among the structural conditions favorable for the formation of falls are the following: (1) harder sedimentary rocks overlying softer ones in a nearly horizontal series, (2) igneous sills or flows interbedded with flat-lying sedimentary strata, (3) successive igneous flows of varying resistance, (4) dikes of igneous rock or any hard layer in other formations, (5) vertical joint planes in massive rocks. Falls may be developed in tributary streams as a result of a rapid deepening of the main valley. Thus, as falls move past the mouth of a tributary, a smaller falls begins in the minor stream. Also, after glaciation it is found that the ice has eroded the trunk valley deeper than the side valleys, and the latter are left as hanging valleys with falls at the junction.

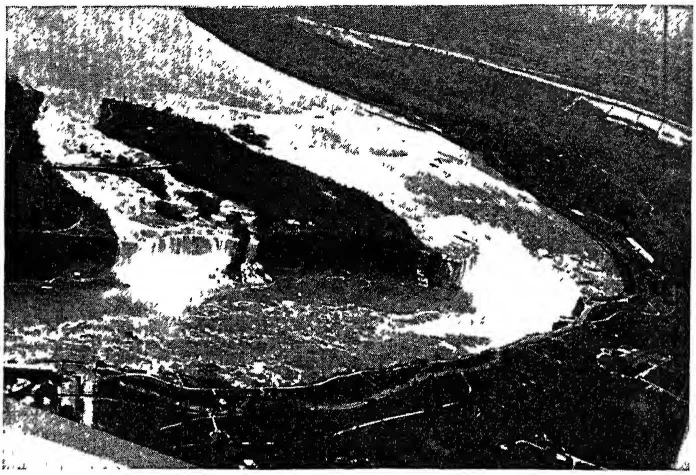

FIG. 128.—Niagara Falls as viewed from the air. (*Photograph by Galloway.*)

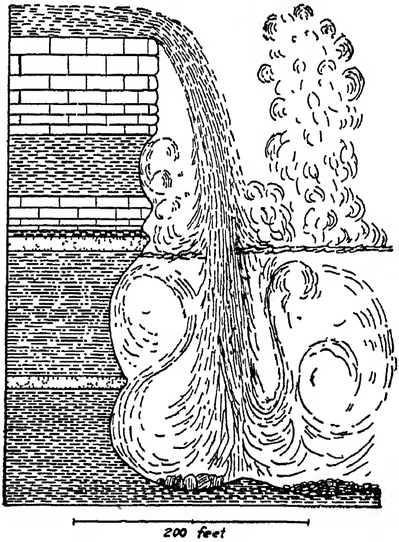

FIG. 129.—Section of Horseshoe Falls, Niagara, showing the structure and succession of strata. About half the drawing is below water. (*After Gilbert, U. S. Geol. Survey.*)

Niagara Falls (Fig. 128) offer a magnificent example of an escarpment formed in nearly horizontal beds where a resistant rock caps notably weaker strata. This cataract plunges about 160 feet over a brink-making limestone about 80 feet thick beneath which is a very soft shale formation. The falls are divided into two parts by an island in the stream channel. The American Falls has a frontage of about 1,060 feet, and the Horseshoe or Canadian Falls has a curved frontage of nearly

2,800 feet. Approximately 500,000 tons of water per minute, or nearly 94 per cent of the water of the river, passes over the Horseshoe Falls. This great body of water has carved out a gorge 200 feet deep beneath the level of the water below the falls, so that the total depth of the gorge from the rim to the bottom of the water is about 360 feet (Fig. 129). As the swirling water back of the falls loosens the soft, shaley formation, it removes it piecemeal and undermines the capping limestone, until finally the limestone cap remains as an inadequately supported overhanging ledge from which large masses of rock plunge into the pool at the bottom of the falls. This process of undercutting is termed *sapping*. Thus, foot by foot, the escarpment has receded up the river, leaving the deep gorge that marks its course from Lewistown on Lake Ontario to its present position, a distance of about 7 miles (Figs. 130, 131).

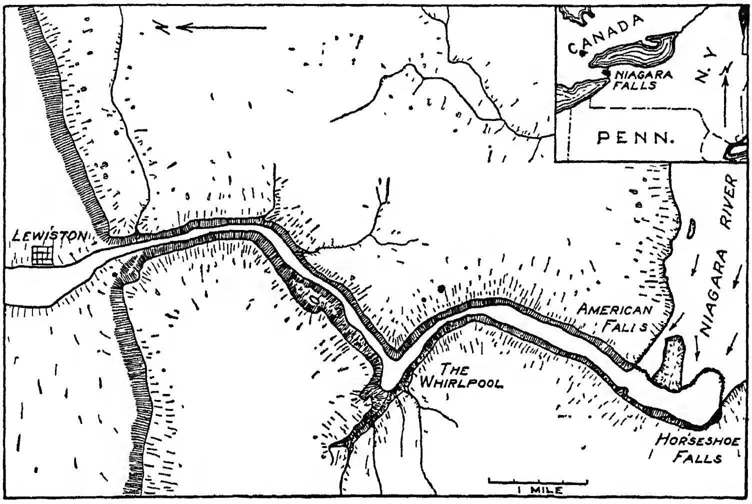

Fig. 130.—Sketch map of the Niagara Gorge showing the course followed by the recession of the falls.

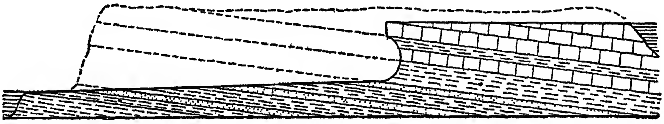

Fig. 131.—Cross section showing the structure of the rocks through which the Niagara Falls have receded. The vertical scale and angle of dip are exaggerated.

The rate of recession of Niagara for an extended period is difficult to determine. A series of factors of unknown value enter into the problem, such as the variation of the river volume and the increase in thickness, toward the south, of the capping limestone rock. Measurements have been made from time to time, and by a comparison of records it has been ascertained that the average rate of recession for 42 years was approximately 3 feet a year (Fig. 132). The Horseshoe Falls is now receding much more rapidly than the American Falls, because of its greater volume of water. Since the American Falls receives scarcely 6 per cent of the stream's water, it is receding only a few inches per year. If the falls had receded 3 feet per year, 12,320 years would have been required for it to have receded 7 miles.

Other Waterfalls.—Along the north shore of Lake Superior many high-gradient streams have cut steep gorges into the Keweenawan lava flows and exposed conglomerate beds separating the successive layers of basaltic rock. At many places the conglomerates or softer amygdaloidal flows are more easily eroded than the overlying massive basalts, and numerous waterfalls result.

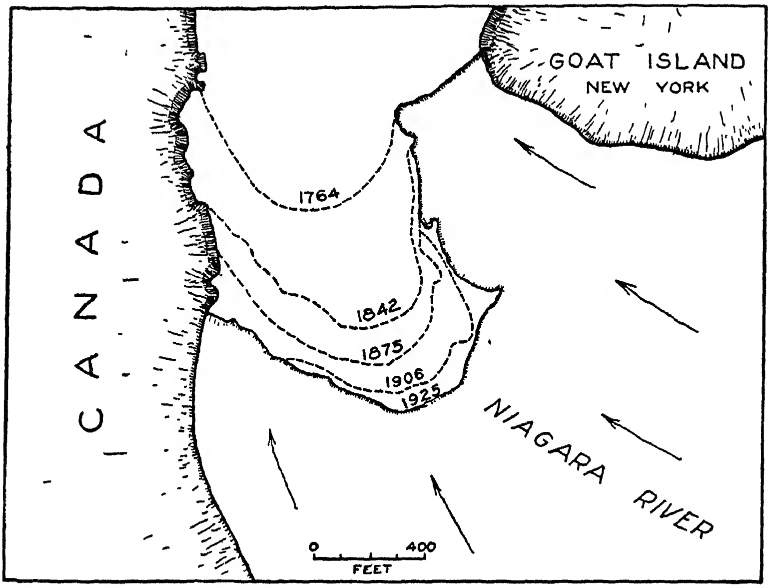

FIG. 132.—Map showing the rate of recession of Horseshoe Falls at Niagara from 1764 to 1925. (*After Niagara Power Co. map.*)

At Yellowstone Falls an irregular mass of more resistant igneous rock crosses weaker rocks in which the deep canyon of the Yellowstone River is excavated. Such falls will not recede a very great distance, for gradually the resistant mass will be cut away, and, as it is worn down, the softer formations upstream will be eroded to the gradient of the river; the crest of the falls will sink, and the falls eventually will disappear or become rapids.

In mountains that have been glaciated, the trunk-stream valley commonly is eroded deeper than tributary valleys because the glacier of the trunk stream is much thicker than the glaciers of the tributary streams. When the ice has melted, these glacial valleys again become watercourses, and the water of a tributary stream falls into the main stream over a steep precipice. Many of the highest waterfalls of the world were formed this way. Yosemite Falls in California (Fig. 133) plunges over a granite cliff into the Merced River valley with an initial drop of 1,430 feet. It then cascades for about 800 feet over a jagged surface with a steep slope and finally plunges 320 feet more over a vertical cliff to the flood plain of the river. Prior to glaciation, the Merced valley possessed a typical V-shaped profile, but during the great ice age a glacier slowly ground its way through the valley, deepened it greatly, and shaped its sides into vertical cliffs over which the tributaries now discharge as falls.

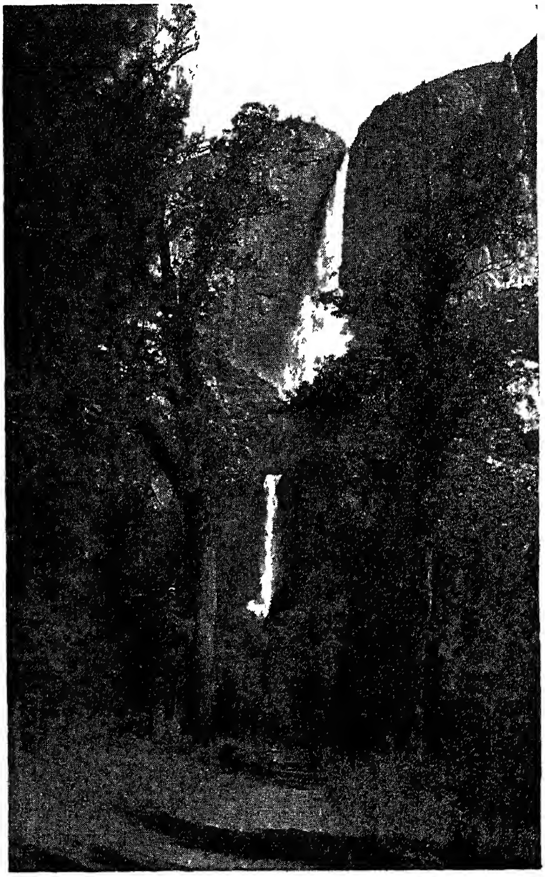

Fig. 133.—Yosemite Falls, in Yosemite National Park, California. A tributary falls discharging from a hanging valley. (*Courtesy Southern Pacific R. R.*)

Vertical joints in massive rocks influence stream erosion much as bedding planes in inclined strata (Fig. 134). Such joint planes are widened by erosion, and large blocks of rock may be removed from the stream bed. As the blocks are removed, a vertical cliff may be developed over which a falls is initiated. In New York near Ithaca, and also at Trenton, falls are developed in jointed limestone.

In mountain ranges where zones of hard rocks cross stream beds, such rocks are eroded more slowly than the softer rocks below them. Vertical escarpments may not be developed, and under such conditions the water leaps from ledge to ledge in a series of sparkling cascades, which may grade imperceptibly into rapids. Along the east margin of the Appalachian Mountains, roughly paralleling the Atlantic Coast, the Piedmont plateau area is bordered by a coastal plain composed of sands and clays not yet indurated into solid rock strata. Streams, flowing from the Appalachian highlands toward the Atlantic, pass from the hard crystalline rocks to the soft unconsolidated sediments nearer the coast, where they cut more rapidly and develop rapids and falls along the line of contact of the two different types of rocks. Because of the great

FIG. 134.—Giant Stairway Falls, Paradise Creek, Alberta. The fall escarpment consists of a series of low, step-like falls. The treads and risers of the steps are joints. (*Courtesy Can. Geol. Survey.*)

number of cascades along a relatively narrow zone, the region is referred to as the *fall zone* (Fig. 135).

Under exceptional conditions, falls may originate by damming due to the deposition of material in the stream bed by processes other than stream erosion. Landslides, glacial deposits, or lava flows (Fig. 136) may form temporary rapids.

FIG. 135.—Map showing the location of the fall zone in southeast United States. Streams flow from the area of crystalline and consolidated rocks of the Piedmont upland west of the zone, to the unconsolidated sedimentary rocks of the coastal plain. Numerous waterfalls and rapids have been developed due to differential stream erosion.

Potholes.—Where rapidly flowing streams produce eddying currents there is a tendency toward the concentration of the energy of the stream at certain places along the channel. If water that is carrying silt is given a rotary motion by the eddies at such points, it tends to grind out round or kettle-shaped excavations in the bedrock of the valley floor (Fig. 137). These are *potholes,* or *giant's cauldrons.* Once a rounded depression is started, the swirling currents in the excavation have their velocities increased during periods of high water, and the rate of deepening is accelerated. Certain potholes are spiral shaped and have a larger diameter near the bottom than at the top. They may be formed either in massive igneous rocks such as basalt or in granite or in softer sedi-

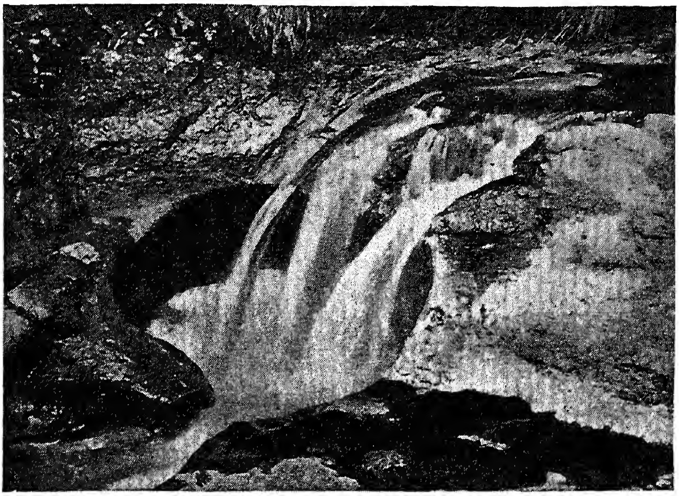

FIG. 137.—A pothole being scoured out by stream action. (*After R. S. Tarr.*)

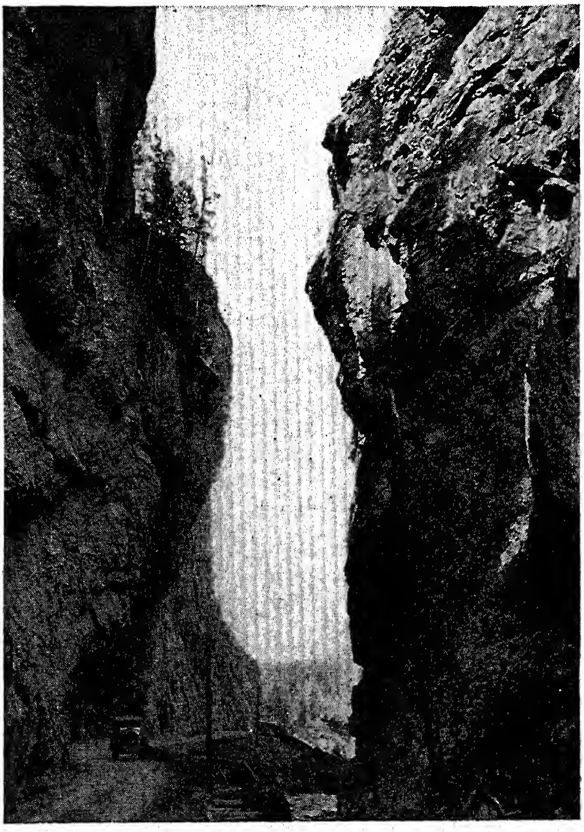

FIG. 138.—Sinclair Canyon, Kootenay National Park, British Columbia. (*Courtesy Can. Geol. Survey.*)

plateau 6,000 to 8,000 feet above sea level. It trenches the high plateau of northern Arizona with a colossal canyon 220 miles long and more than a mile deep. Where it reaches this great depth, the total width from rim to rim is from 8 to 12 miles, but its width at the bottom is only slightly greater than that of the stream. If the slopes of the canyon were uniform, it would have an angle of less than 15 degrees, but the inequalities of hardness of the sedimentary strata have produced gigantic steplike slopes, or rock terraces, some of the step faces of which drop vertically for more than a thousand feet. The upper series of rocks in which the canyon is cut is composed of flat-lying beds of limestones, sandstones,

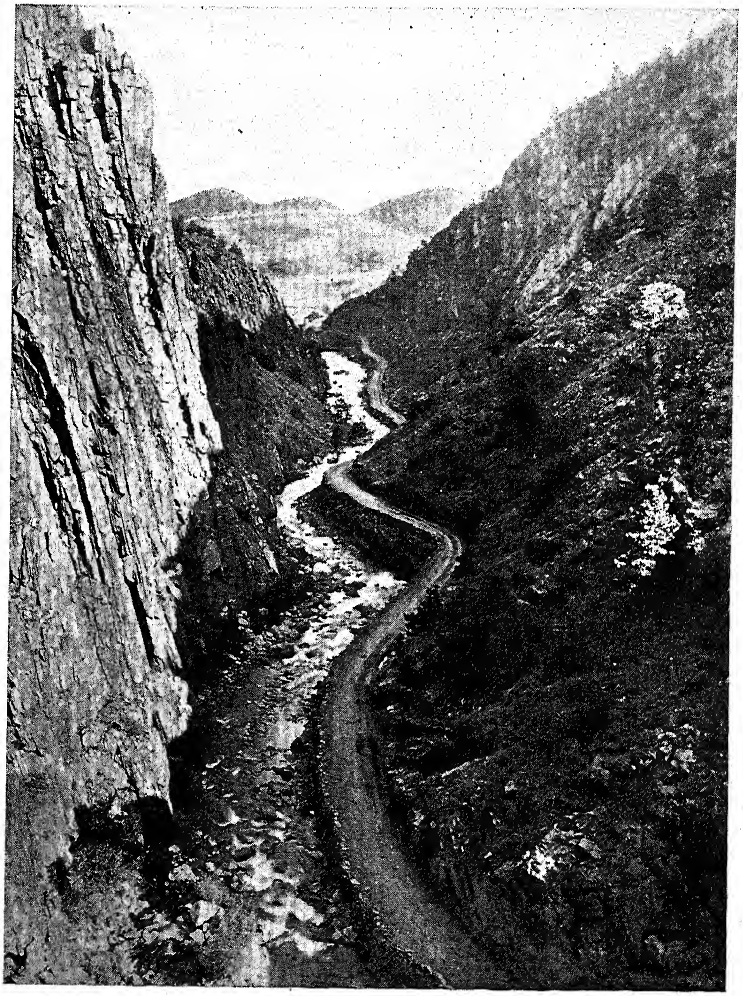

FIG. 139.—Big Thompson Canyon, north of Denver, Colorado. A steep-walled canyon cut through steeply dipping sedimentary and metamorphic rocks. (*Courtesy Denver Tourist Bureau.*)

and shales (Fig. 85). Beneath this the stream has disclosed ancient crystalline schists and massive igneous formations.

Natural Bridges.—An arch of rock across a valley is an exceptional and striking feature of the topography (Figs. 140, 141). There are

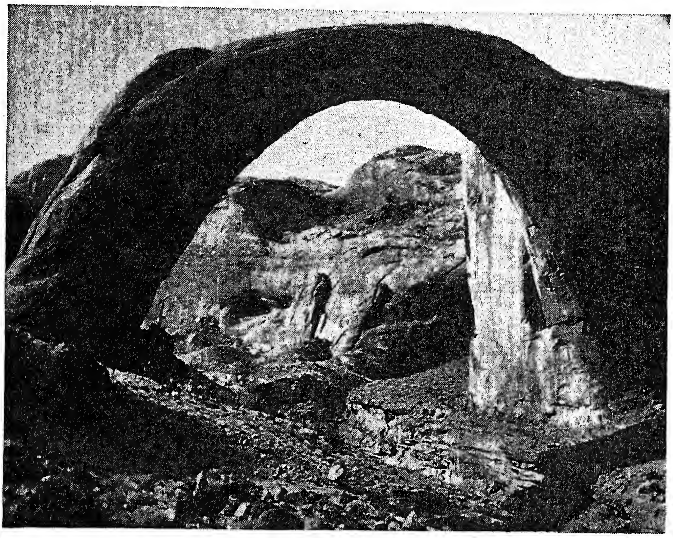

FIG. 140.—Rainbow Natural Bridge, southern Utah. Formed as shown in Fig. 141. (*Courtesy Santa Fe R. R.*)

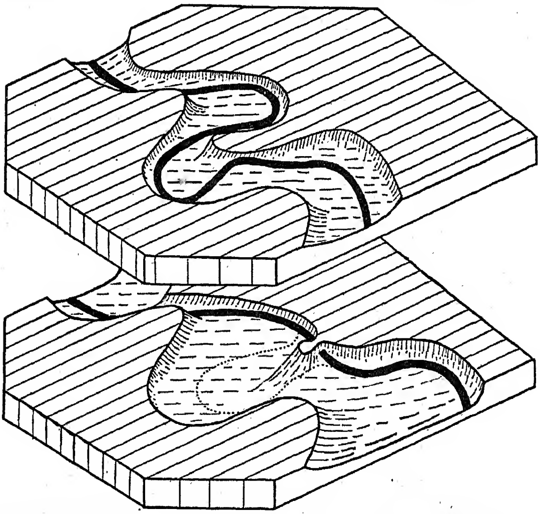

FIG. 141.—Diagram showing the formation of a natural bridge by the perforation of the upland in the loop of a meander.

numerous ways in which such natural bridges may be formed (see page 84). (1) Where a surface stream disappears into the joints of the bedrock of a valley it may flow under ground for some distance and then reappear at the surface. As erosion continues, a "valley" is excavated beneath the surface, and the rocks will span the valley as a bridge. (2) If the rock of a stream bed is jointed above a waterfall, some of the

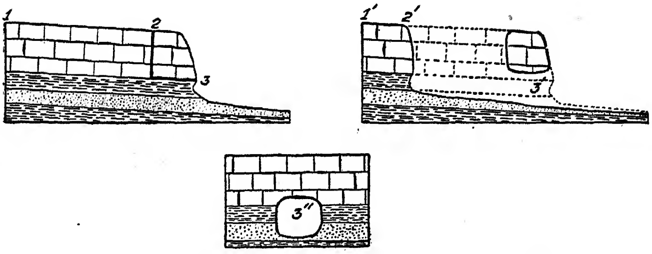

Fig. 142.—Diagrams illustrating the formation of a natural bridge by the enlargement of a fracture (2) across a stream bed upstream from a waterfall. The solution of the limestone produces a tunnel at 3, which in time becomes sufficiently large to carry the entire volume of the stream (3′). The valley is then spanned by a natural bridge, as shown in lower diagram.

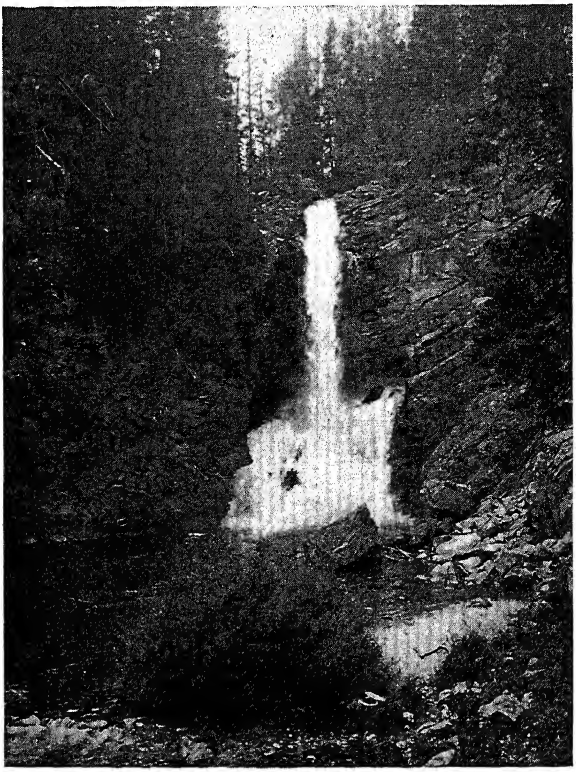

Fig. 143.—Trick Falls, Glacier National Park. A natural bridge in early stage of formation. Some of the water flows through a joint system not seen and reissues near the foot of the falls to rejoin the part of the stream that flows over the falls. (*Photograph by Miss Ethel M. Rodgers.*)

water may descend through a joint and then follow a bedding plane until it issues into the main channel below the falls or back of the curtain of water forming the falls (Figs. 142, 143, 144). The joint is slowly enlarged until the channel becomes large enough to accommodate all of the water, and the mass of rock from the escarpment that produced the waterfalls upstream to the position of the vertical joint remains as a natural bridge. In Two Medicine River in northwestern Montana this process is now in progress, and a natural bridge is partially developed. (3) Where ancient peneplanes have been rejuvenated, many of the meandering streams become deeply intrenched, and the lateral swinging of a stream against the cliff-like walls undercut the neck of a meander at the level of the water. This process continues from both sides until a hole is cut through and the stream flows through the perforation, leaving an arch of rock as a bridge over the stream (Figs. 140, 141). The famous Rainbow Natural

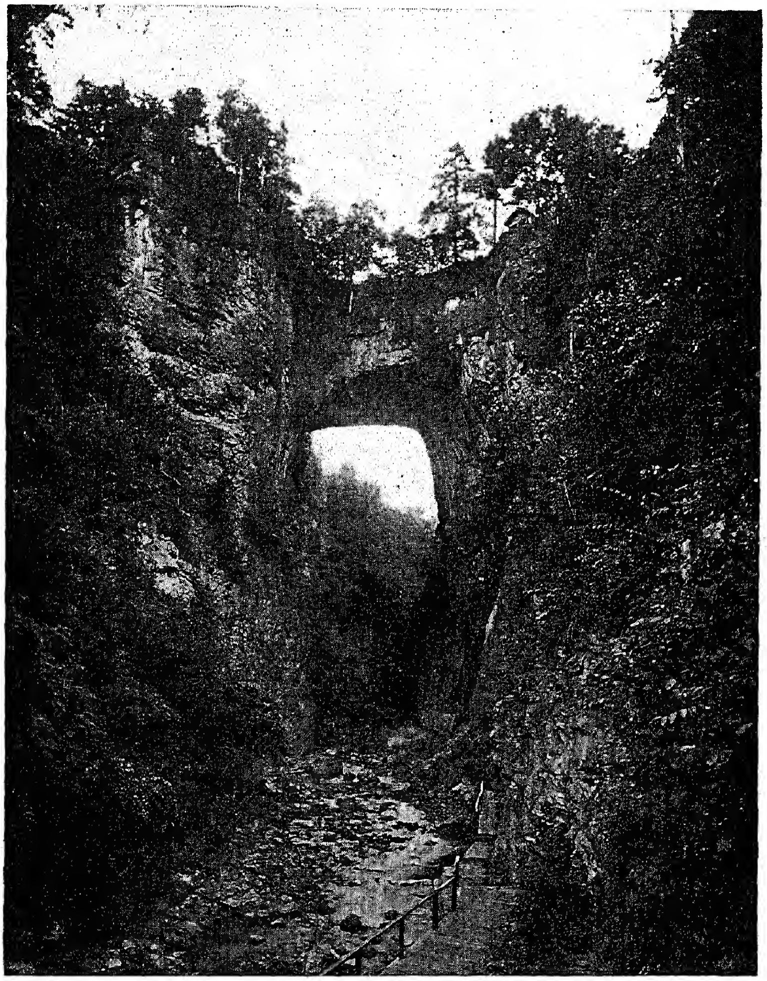

FIG. 144.—Natural Bridge, Virginia. (*Photograph by Walcott, U. S. Geol. Survey.*)

Bridge in San Juan County, Utah, is an example of a bridge so formed. (4) Where the tributary streams are extended into steeply sloping divides by headward erosion the watershed may be reduced to a very narrow ridge. If the divide is capped by a resistant formation overlying soft sand or shale, the divide may be perforated under the more massive cap rock. (5) In the petrified forest of Adamana, Arizona, a silicified log has been undermined by stream erosion until a very unusual natural bridge spans the valley. The trunk of the petrified log which is about 3 feet in diameter lies diagonally across a canyon about 30 feet wide and 200 feet deep. The rock in which the log was imbedded is poorly cemented and therefore easily eroded. (6) In the volcanic areas of the West many natural bridges have been formed by the incomplete collapse of the roofs of lava tunnels. (7) Other natural bridges are made by wave erosion, by sand blasting by the wind, and by differential weathering.

STREAM DEPOSITION

Stream transportation is attended by stream deposition. Anything that decreases the transporting power of a stream promotes deposition. Along the course of every river the current frequently is checked, and at such places sediments are deposited. Even the bed of a stream having a relatively high velocity will have part of its surface covered with rock fragments, which the stream was forced to drop because of changes in the transporting power of its currents. Since conditions favoring transportation vary from time to time and from place to place, sediments derived from the land generally are not all carried directly to the sea, but some may be deposited to form definite features in the topography as a region passes through a cycle of erosion.

Factors Causing Deposition. 1. *Diminished Velocity.*—A slight diminution of the rate of flow of a loaded stream will initiate deposition. A loss of velocity may be brought about (*a*) by a decrease of slope or gradient of the stream bed; (*b*) by a decrease in volume of the stream; (*c*) by a change in the configuration of the valley; (*d*) by encountering obstructions such as heaps of residual boulders formed from more resistant dikes or beds that cross the valley or by temporary dams formed by floating trees or rafts of logs; and (*e*) by flowing into a body of quiet water such as a lake, estuary, or bay.

2. *Diminished Volume.*—Since the volume of water affects the carrying power of a stream directly, variations in volume cause streams alternately to aggrade and degrade their beds. Many streams have seasonal high-water stages, during which more and coarser sediments are carried, and later these are deposited when the amount of rainfall decreases. Diminished volume may result also in other ways. Some of the water of the stream may sink into the earth where the stream flows

through a dry region. In such regions the bottom of the valley does not reach the ground-water level, and some of the water of the stream is absorbed by the soil and rocks through which it flows. Where a stream breaks up into a number of distributaries, the volume of each branch is less than that of the original stream. If, as a result of division, the stream drops much of its load in its own channel, it then flows as small streamlets through and over its own sediments.

Place and Manner of Deposition.—Stream sediments or alluvium may accumulate in a number of different places such as (1) at the foot of mountain slopes, (2) in stream channels, (3) on river banks and flood plains, (4) at debouchures.

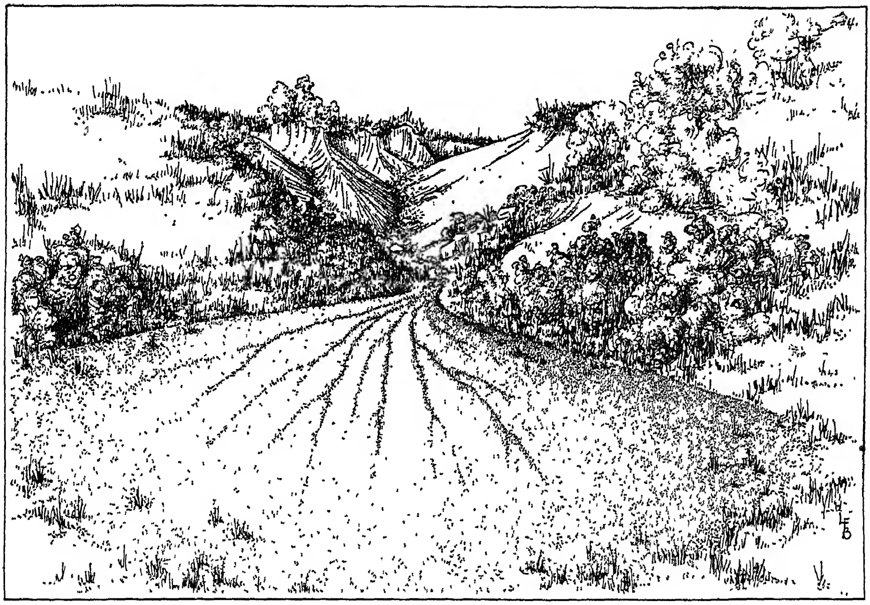

FIG. 145.—Sketch of a young stream valley with an alluvial fan.

Alluvial Fans or Cones.—Where a stream descending a steep slope issues from the mountains on a plain or in a wide valley, its velocity suddenly is diminished, and a large part of its load is deposited and spread out in the form of a fan-shaped heap at the opening of the ravine or gully through which the stream flows (Fig. 145). As the deposit thickens, its thickness becomes greatest at the mouth of the steep valley, and a cone-shaped structure is developed. The slope of the cone's surface varies with the size and velocity of the stream and the character of the transported sediment. Where streams discharge on a plain near each other, their fans may coalesce and form a continuous sheet of aggraded sediments along the base of the mountain range and eventually build up an alluvial piedmont plain (Fig. 146).

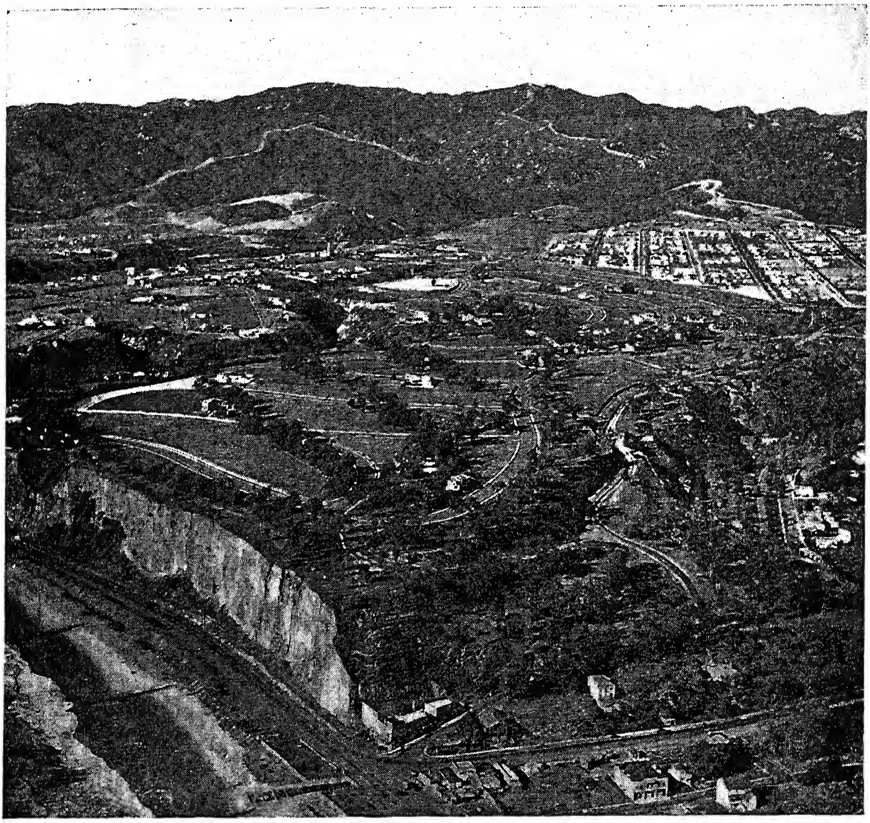

FIG. 146.—Dissected piedmont fans. Huntington Palisades, near Santa Monica, California. (*Courtesy Spence Air Photos.*)

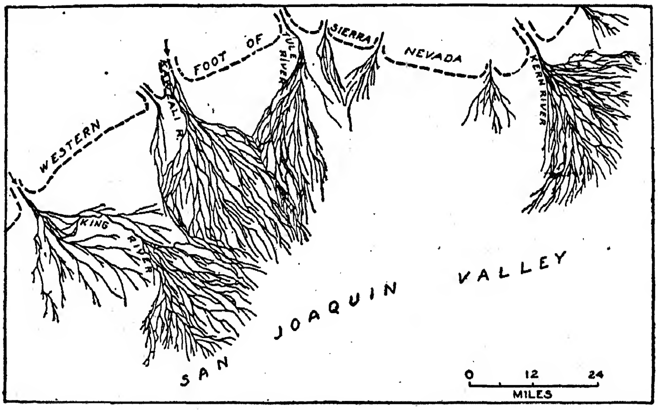

FIG. 147.—Sketch map showing the distributaries and the braided patterns of the streams from the Sierra Nevada on their alluvial fans in the San Joaquin valley in California. The water of these streams enters the porous sediments that constitute the growing alluvial fans. (*After W. D. Johnson.*)

Conditions favorable to the formation of fans or cones are not entirely topographic. Climatic variations also may be a factor. An abrupt change of slope was the chief factor in the formation of the fans at the western base of the Sierra Nevada Mountains in California (Fig. 147) and at the eastern base of the Front Range of the Rocky Mountains. Huge fans have been built up also at the base of the Himalaya Mountains in India and at the base of the Andes in Argentina and at many other places.

Deposits in Stream Channels.—The deposition of sediments in a stream's channel is characteristically shown by the accumulation of alluvium at the inner side of each sharp bend of a stream's course. The main current makes a rapid sweep along the outer bank, and undercurrents pass across to the inner side of the curve and deposit parts of their loads. Deposits so formed, when exposed during the low-water stage of a stream, are the familiar sand banks or pebble beaches of streams.

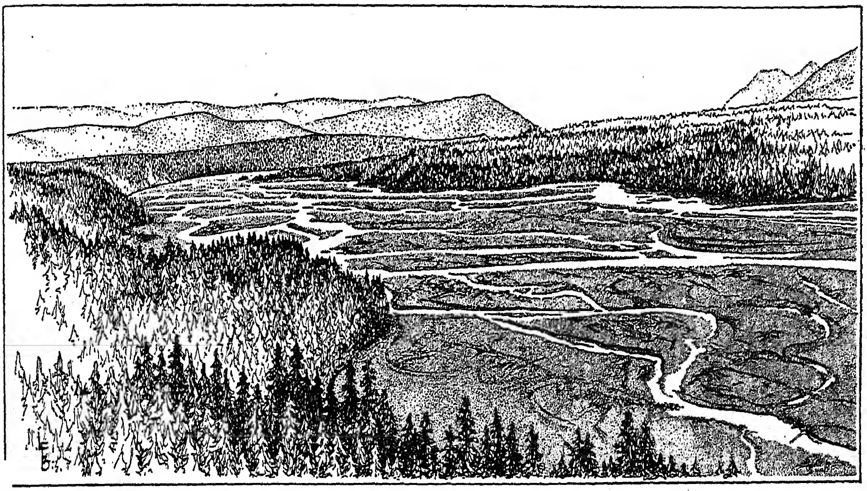
FIG. 148.—An overloaded stream, Matanuska River, Alaska. (*Sketched from photograph by Mendenhall, U. S. Geol. Survey.*)

Braided Streams.—In streams that are heavily loaded (Fig. 148), especially in those subject to loss of volume in semiarid regions, sediment is being deposited continually, and the bed is steadily rising until eventually numerous sand bars deflect the currents, and the river flows not in a single channel but in many anastomosing streamlets which together with the sand bars that separate them are continually shifting. A river which, owing to deposition, is split into many branching and reuniting channels is a *braided stream.* Platte River in Nebraska is a braided stream flowing in a broad alluvial valley nearly a mile wide. During most of the year a small volume of water finds its way in a tortuous course through a series of interlacing streamlets whose positions shift at every flood.

Scour and Fill.—When the volume and velocity of an aggrading stream are suddenly increased, as in time of flood, it digs new channels in the sediments on its floor (Fig. 149); and when the flood subsides, they are again filled. Such alternate filling and excavating is referred to as *scour and fill.* One of the most striking examples of a stream that transports its sediment in this manner is the lower portion of the Missouri River. During periods of high velocity, scouring reaches a depth of about 80 feet in the vicinity of Nebraska City, and 25 miles upstream from Omaha a fill of about 40 feet is cut to bedrock during seasons of flood. The products of such excavation are moved downstream and eventually, after many periods of rest, reach the sea.

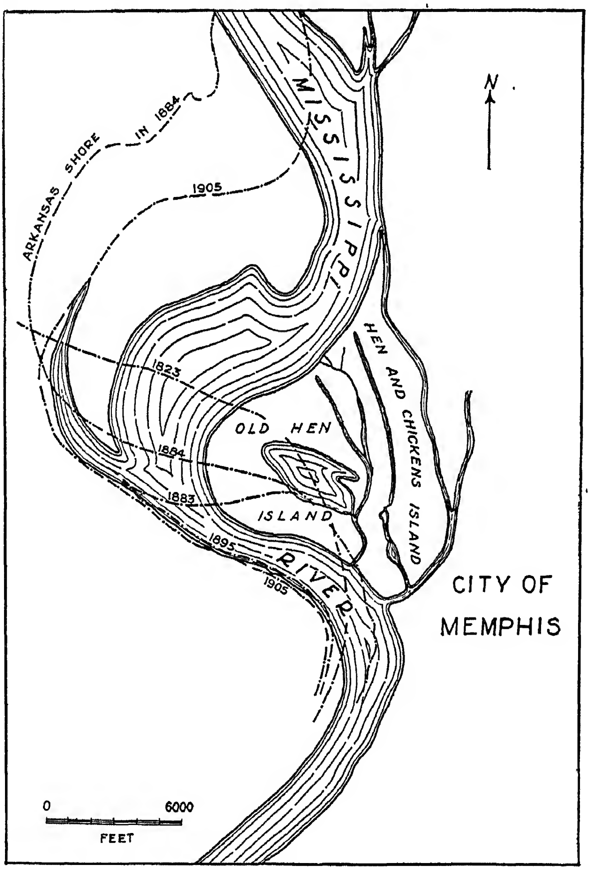

FIG. 149.—Sketch map showing the shifting of the bank of the Mississippi River channel in the region of Memphis, Tennessee. (*After M. J. Munn, Tenn. Geol. Survey.*)

Deposition on River Banks and Flood Plains.—In time of flood, when the volume of a stream is high, fine silt, mud, and sand are laid down on the level tract or flood plain over which the river spreads. During each high-water stage, the bed of alluvium becomes thicker, and the height of the flood plain is increased until the deepening of the main channel by erosion makes the height of the flood plain above the normal stream so great that the plain is no longer overspread by the river, except in times of very high floods. The part of the flood plain of the Mississippi River from Cairo, Illinois, at the junction of the Ohio River, to the Gulf of Mexico varies from 30 to 60 miles in width and is approximately 600 miles long. Most flood plains are bounded on either side by relatively steep slopes. These slopes may be sufficiently steep to form bluffs, especially if the valley has been widened by lateral cutting in resistant rocks. In

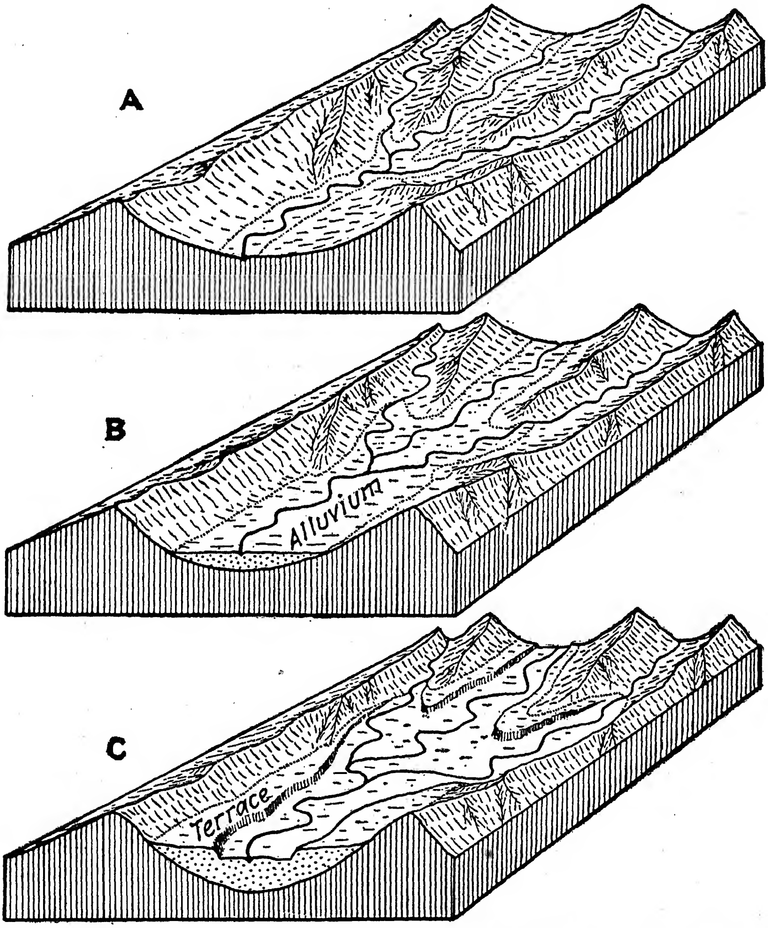

Fig. 150.—Diagrams showing the development of stream terraces. *A*, stream degradation; *B*, aggradation with the development of flood plain; *C*, terraces carved from the flood plain by river, which erodes part of its earlier deposits. (*Based on figure by Emerson.*)

some cases where the rocks bordering a valley are weak and easily eroded, the slopes are so gentle that it is difficult to detect where the flood plain ends and the valley sides begin.

Flood-plain Terraces.—A stream that has aggraded its valley to a considerable depth may excavate part of the deposits previously laid down. The alluvial sediments are carved into one or more terraces or narrow plains and flats that fringe the sides of the valley. Such flat-topped stream terraces are the remnants of former flood plains, below which the streams that made them have cut their channels to develop new flood plains at lower levels (Fig. 150).

The chief factors that aid the development of stream terraces are (1) uplift or rejuvenation of the stream, (2) partial loss of load and renewed ability to erode farther downstream, (3) failure of supply of sediments in the upper stream course, (4) exchange of a small amount of coarse for a large amount of fine sediment, (5) elimination of meanders and consequent increase of velocity, and (6) increase in volume of the stream due to piracy or to other causes.

Terraces are a normal feature of the history of any stream. They appear first in the lower or older part of the valley and are gradually extended upstream. Where traced upstream or downstream they are found to pass into the flood plain. It is rare that terraces of the same age are equally developed along both sides of a stream at the same time, for the new or lowered channel is likely to lie near one side of the old flood plain. Even if terraces are formed on both sides, erosion is constantly attacking them and may destroy them; and the older the terrace the smaller the remnants become until finally it disappears.

Natural Levees.—During floods the whole flood plain may be covered with waters flowing seaward. In this wide expanse of water the current is most rapid along the axis of the river channel, where the water is the deepest. Along the margin of the channel, where the rapid currents come in contact with the slowly moving water of the flood plain, the velocity suddenly is checked, and the currents drop all but fine sediments carried in suspension. In this way the flood-plain deposits are built up highest on the immediate border of the channel and slope gradually toward the valley sides. These embankments of aggraded material resemble the levees constructed by man in his attempt to confine a stream to a narrow channel, and they are known as natural levees (Figs. 151, 152).

The levees are low ridges seldom more than a few feet higher than the back-land toward which they descend with a slope so slight that the region appears flat. Some levees however, are so high that during flood time they stand out as long, low islands with the main channel of the river on one side and the flood waters of the backland and uplands on the other side. During moderately high water, natural levees serve as a protection to the river flats, as they retain the waters in a definite channel. However, during unusually high floods a river may break through its embankments and flood the lowlands beyond with numerous streams.

Along many rivers the levees are so high that tributaries flow parallel to the main stream for considerable distances before they find a place to join the main stream. The Yazoo River parallels the main channel of the Mississippi River for about 200 miles.

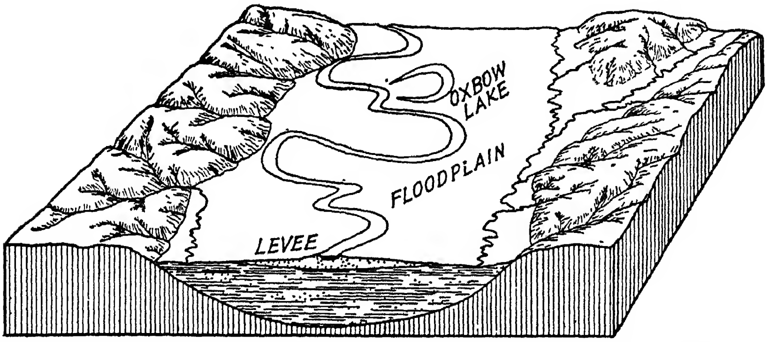

Fig. 151.—Diagram of the flood plain and natural levees of a large river. Note the oxbow lake and marginal streams.

Many cities are built on flood plains because they offer flat ground for buildings near water transportation. Such cities are almost certain to be inundated during high water. Disastrous floods occur nearly every year along the Mississippi River and its tributaries. Since certain cities are built on plains formed at flood times of the streams (Fig. 152), frequent inundation should be expected.

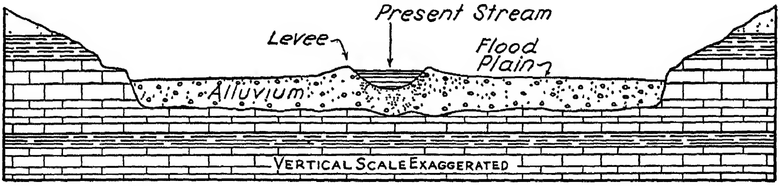

Fig. 152.—Diagram showing flood plain of a river with levees. If a city is built on such a plain, frequent inundation is to be expected.

Deltas.—Where rivers laden with sediments flow into a body of quiet water such as a lake or a bay or into the sea, the velocity suddenly is checked, and rapid deposition of sediments follows. Where the shore currents are of insufficient strength to transport the load brought in by the river, or where coastal configuration protects the mouth of the stream from rapid tidal currents, the débris which the river brings settles near its mouth and builds up a delta, named from the Greek letter Δ, the shape of which it somewhat resembles (Fig. 153). Where a river discharges into a sea, the silty, fresh water tends to float on the heavier salt water until the fresh and the saline waters mix. Where fresh water, with fine silt in suspension, mixes with sea water, the salts of sea water cause the silt to be deposited.

Since waves, tides, and currents are weaker in lakes than in the sea, deltas are more common in lakes. However, the larger deltas are built in the sea, for there the largest rivers discharge. Some streams build no deltas because of the lack of sediments. The Niagara River as it flows into Lake Ontario is so free from silt that no sediments are being deposited.

Growth of Deltas.—A delta consists of successive layers of débris brought down from the land and spread out over a fan-shaped area on

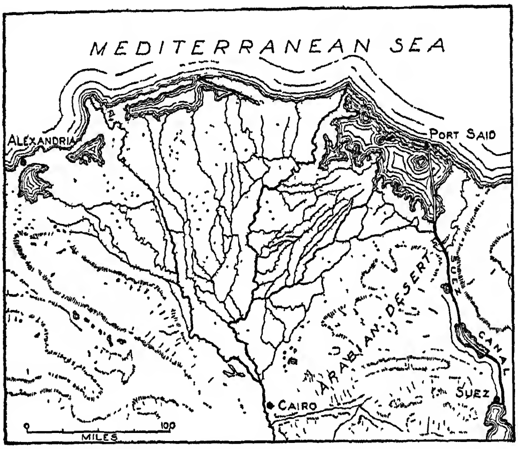

Fig. 153.—Outline map of the Nile delta. (*After Kayser.*)

the bottom of the basin at the mouth of the river. Where the stream current reaches quiet water, the bulk of its coarser load is dropped, and the finer material is carried farther out. The accumulating sediments tend to reduce the gradient of the stream so that it aggrades rapidly, and soon it begins to break up into distributaries which wind to and fro over the newly formed alluvial land (Fig. 154). Deposition continues in these currents until many of the channels are completely choked and others are opened; later these also are choked and abandoned. In this way many partially filled areas remain as lakes in the delta, and coalescing distributaries inclose islands. The main channels build their accumulations of coarse débris farther and farther out into the sea, and the minor distributaries assist by adding their products of aggradation until the deposits are built up to and above sea level.

The rate of growth of deltas varies with the size and velocity of the rivers and the geologic nature of the drainage basins. The Mississippi

River is pushing the embankments of its chief distributaries into the gulf at a rate of about a mile in 16 years (Fig. 155). The northern portion of the Adriatic Sea is being filled so rapidly by the Po and other streams that cities originally built on the coast line are now far from the sea. Adria, formerly a port, is now 14 miles inland; and on other parts of the coast line, zones 20 miles wide have been built up within the past 1,800 years. The Tiber River, yellow in color because of the abundance of sediment which it carries, is adding to the coast line around its mouth at the rate of about one mile a century. At the mouth of the Danube a

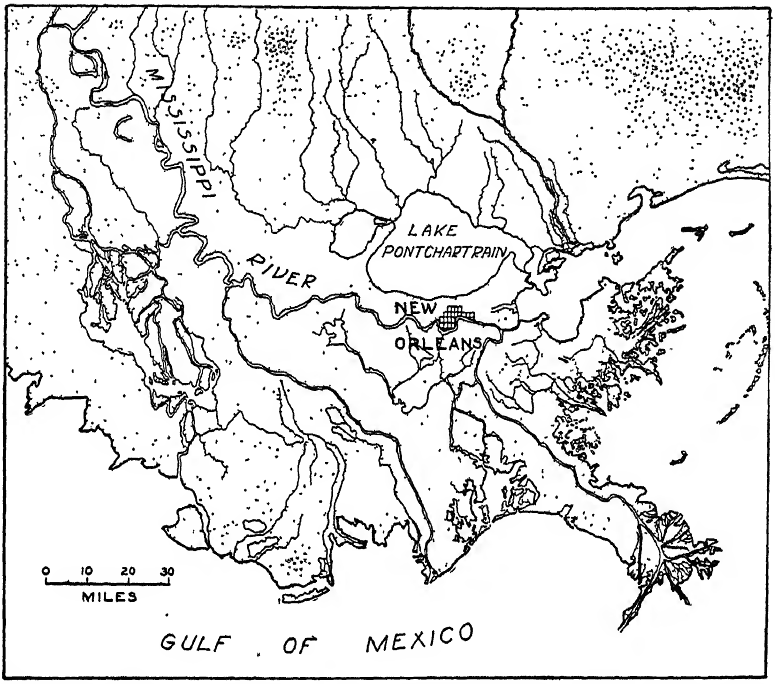

FIG. 154.—Map of part of the delta of the Mississippi River. (*Drawn from recent maps and aerial photographs by Trowbridge.*)

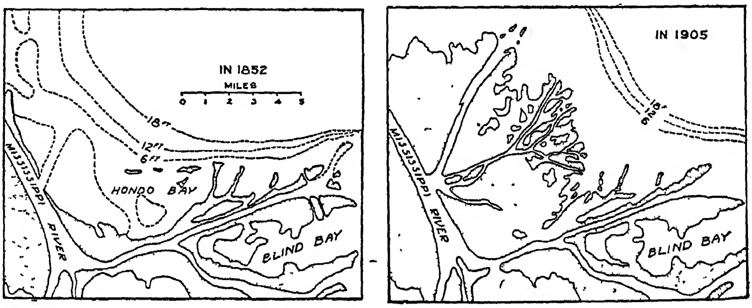

FIG. 155.—Part of the Mississippi River delta showing the amount of growth during a period of 53 years. (*After G. R. Putnam.*)

great delta is growing into the Black Sea. At the major outlets the water is shallowing so fast that the lines of soundings of 6 and 36 feet deep are advancing into the sea nearly 400 feet per year.

The Ganges-Brahmaputra delta has an area of approximately 60,000 square miles with an apex 200 miles inland. The Mississippi delta began to form north of Cairo, Illinois, and delta-like deposits are found southward for 600 miles. Its width varies from 30 to 60 miles. The total area including its northerly extension is over 30,000 square miles. Only about one-third of this area has been formed during recent geologic time. Most of Holland represents the combined deltas of the Rhine, the Maas, and the Schelde rivers.

Structure of Deltas.—The structure of a delta deposit is essentially the same at the mouth of every large river. A section shows nearly horizontal beds of fine silt at the bottom. Because of their position these beds are termed the *bottomset* beds. Above them is found a series of more steeply inclined strata of coarser sand and gravel which represents the heavier load dropped by the river currents as they encountered the quiet waters of the coast. The slopes of these beds, called the *foreset* beds, approach the angle of repose of the material of which they are composed. They, in turn, are capped by nearly horizontal strata, the *topset* beds, which represent the last deposits of the river as the distributaries aggrade the alluvial flats of the growing delta (Fig. 156).

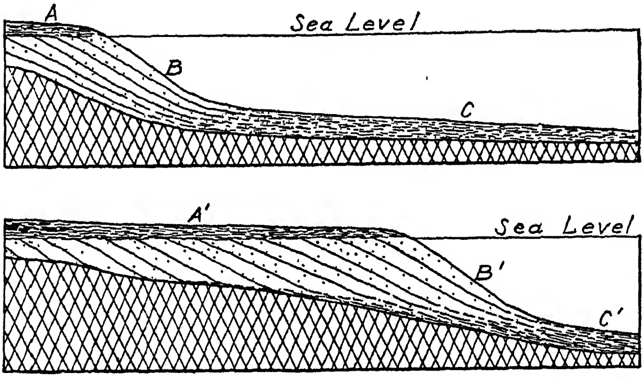

FIG. 156.—Diagram illustrating the development of a delta: A and A', topset beds; B and B', foreset beds; C and C', bottomset beds.

A delta consists of two parts: (1) a seaward margin, or submerged part, resting upon the marine continental shelf; and (2) a landward part that is not submerged but is covered with the subaerial and fresh-water sediments. The land portion gradually is built out over the submerged portion, which in turn steadily is built out into the deeper water. Because of these relations, oceanic deltas have marine beds that grade upward and landward into fresh-water sediments.

Subsidence of Deltas.—Subsidence of large deltas has been noted at the mouths of many rivers. In general, aggradation has kept pace with the sinking, but in some delta deposits marine limestones and shales are interbedded with fresh-water sediments and old soils. From the structural relations of the sediments it is evident that at times subsidence gained on upbuilding, and the delta surface again was covered by the sea. Deep borings in various deltas reveal similar successions. At New Orleans driftwood was penetrated at 1,042 feet. Depths from 500 to 800 feet are not uncommon in many deltas now being formed. During earlier geologic times some deltas subsided to depths of more than 10,000 feet as the river-borne sediments accumulated on the surface. In the region that is now the northern Appalachian Mountains area an ancient delta was once deposited, and the Susquehanna River has since excavated its channel through the delta beds, which show a thickness of 13,000 feet.

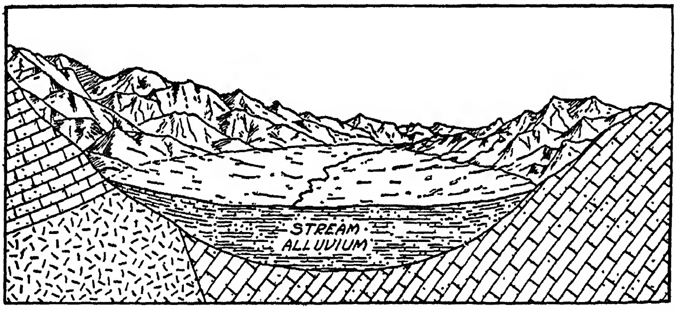

Fig. 157.—An aggraded structural valley in a semiarid region. (*After U. S. Geol. Survey.*)

Stream Deposition in Arid Regions.—In arid regions the rainfall is so slight that the run-off is not able to transport all the sediments brought into the larger valleys by the rain wash. For this reason a large proportion of the material derived from the erosion of the bordering hills or mountains where the rainfall may be considerable is deposited in the valleys, where rainfall is much less, with the result that sediments accumulate to a depth of many hundred feet (Fig. 157). Desert-valley filling characteristically is shown in the smaller intermontane valleys of Arizona and California. There many of the valley walls are angular mountain slopes with narrow, gorge-like tributary valleys leading back into the mountains. A huge alluvial fan or cone is formed at the mouth of each tributary, and where the gradients are high, many of the fans are composed of boulders and large stones, transported by the torrents produced by cloudbursts of only a few minutes' duration that sweep over such areas at rare intervals.

In regions where the major valleys are wider, the outer margins of the fans are composed of fine sand and silt, and here wind work becomes active, and sand-dune areas are developed. In many of the valley bottoms the wind is a more effective agent of transportation than running water. Occasionally, however, when the rainfall is heavy in the mountains at the headwaters, a desert valley may have a river that floods the aeolian sediments of its plains for hundreds of miles along its course, so that fluvial and aeolian sediments become interbedded.

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CHAPTER VI

GRADATIONAL WORK OF SNOW AND ICE

Freezing of Ground Water.—Fresh water under ordinary surface conditions forms ice when the temperature reaches 32°F., and as it passes from the liquid to the solid state there is a 9 per cent increase in volume. Since all soils are porous and since nearly all solid rocks near the surface have innumerable joints and fissures that contain moisture, the expansion of water as it freezes performs much work in rock disintegration (Figs. 158, 159). In high mountainous districts and in the polar regions, where there is a marked daily fluctuation of temperature, small, angular fragments are wedged from the surface of exposed cliffs and accumulate as heaps of talus at the base of steep slopes and crags. Along the coast of Greenland and in Antarctica the amount of rock disruption caused by frost is enormous. There some of the snow of the long winters is melted during the summer months, and water fills the joints and fissures of the rocks. The summer nights are sufficiently cold to freeze water, and the resulting ice splits off large blocks from the shore-line cliffs. In mountain valleys with high gradients, talus material creeps under its own weight and frequently forms "rock streams," or "rock glaciers," that move slowly down the valley.

Fig. 158.—Blocks of rock loosened by frost and the expansive force of ice. Yosemite National Park, California. (*After Matthes, U. S. Geol. Survey.*)

In soils the frozen water forms an icy cement which hinders the percolation of ground water as the snow melts in the spring. On hillsides and mountain slopes such a frozen surface becomes a gliding plane in the rock waste, and huge masses of débris slump toward the valley as landslides. In agricultural districts where the soil is composed of bouldery clay, each season's frost carries boulders nearer the surface. The stones are moved by the expansive force of water-soaked clay that yields in the direction of least resistance and, in general, such frost heaving is toward the surface of the earth. When the frost leaves in the spring, the particles of soil sink back to their original positions, but the boulders are prevented from settling by pebbles or clay which fall into cavities that lie below them.

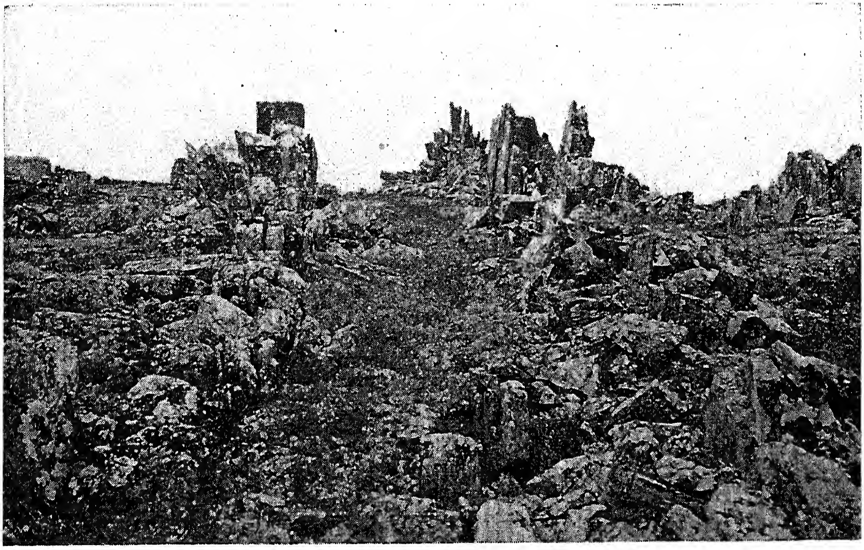

Fig. 159.—Frost-heaved blocks of rocks near Mackay Lake in northwestern Canada. (*Photograph by Hicks, Can. Geol. Survey.*)

Freezing of River Water.—When river water freezes, it exerts a disruptive effect at the sides and bottom of its channel like that which results from the freezing of ground water. Mud, gravel, and boulders become incased in the ice and are pushed downstream or are floated by thick cakes of ice to the banks of the stream or halted by an ice jam across the valley. If a large amount of ice accumulates at a narrow place in the valley, the water is ponded, and deposition of the stream's sediment follows. When the jam eventually breaks, the volume of the stream is greatly increased, and the acceleration of velocity that accompanies it raises the transporting power of the stream immensely. Many of the Canadian rivers flowing into Hudson Bay are dammed at their rapids in this manner. As the ice breaks, in early summer, many of the incased boulders are stranded on the shores, where they remain until they are pushed or floated farther during the next season.

Ice on Lakes and Seas.—A body of quiet water, such as a lake or inland sea, does not commonly freeze at the surface until all of the water from top to bottom is near the freezing temperature. Fresh water is densest at 39°F. and when the surface water reaches this temperature it sinks and is replaced by warmer water. Thus the lakes of average depth freeze over at the surface completely when the temperature is below the freezing point for an extended period. Deep lakes, such as the Great Lakes, freeze near shore lines but do not completely freeze over, even in the coldest winters, since the cold surface water sinks and is replaced by warmer and lighter water from the depths; all of the water in such deep basins does not reach the temperature of greatest density.

If a lake is shallow for some distance from shore, the ice anchors itself to the bottom; and by freezing the water in the sediments on the floor of the basin, the ice covering of the basin becomes continuous with the frozen land at the level of the lake. With fluctuations of temperature the ice expands and contracts, and in contracting during a period of low temperature it forms cracks which are filled with water that congeals to cement the fractures. In this way a continuous sheet of ice is formed again to fit the outlines of the basin. If later the temperature is raised, the ice expands, making the ice cover too large to fit the basin, and it crowds the shore line, exerting an enormous horizontal thrust in all directions. In this way much loose material is crowded ashore, forming, after melting of the ice, walls of sand, gravel, and larger stones that parallel the shore lines. They differ from beaches or bars in that the material is often unsorted, and it slopes steeply toward the basin. Hundreds of glacial lakes in the upper Mississippi valley region have conspicuous ice ramparts formed in this way.

On many lakes under the vegetation of a floating bog, ice is continuous laterally with the ice of the surrounding soil. At such places the expanding ice arches up the soil in a series of ridges parallel to the shore. Many small lakes in the glaciated region of the northern part of the United States are entirely surrounded by wire-grass bogs underlain by thick beds of peat that are saturated with water. The shove of shore ice on such lakes pushes the ice of the surrounding bogs into huge domes and ridges.

Should strong winds arise when the ice on a lake is breaking, large cakes of ice with inclosed pebbles and boulders are driven ashore and pushed out on the beach. If the drift of ice rafts parallels the shore line, erosion of the banks takes place by the ice which tears away the beach materials or smooths them into terrace-like structures.

Sea water does not freeze until it reaches a temperature of 26° to 28°F. The freezing temperature is not so nearly uniform as that of fresh water, because of the varying salinity of the sea. In high latitudes the sea water is all near the freezing point, but after the ice attains a thickness of from 6 to 10 feet, it protects the water below from the intense cold of the polar winters. Where ice forms along shore in the high latitudes, it is at many places over 50 feet thick. Such masses are not the result of the direct freezing of the ocean water but the conversion of snow into ice by the spray from waves and the heaping up of ice along shore by storms. Such ice, commonly referred to as the *ice foot of the shore*, is an important factor in shore-line erosion. It protects the shore from wave action and serves as a raft on which débris, broken from the cliffs by the action of frost during the winter, may gather and be carried out to sea when the ice breaks.

Broken sea ice floats in large cakes known as *floes*. This floe ice frequently is jammed together into ice packs that have very irregular surfaces and stand high above the water, owing to the crowding of the ice cakes. Along the coast of Labrador and in many Arctic bays ice forms on the sea bottom and is spoken of as *ground ice*, or *anchor ice*.

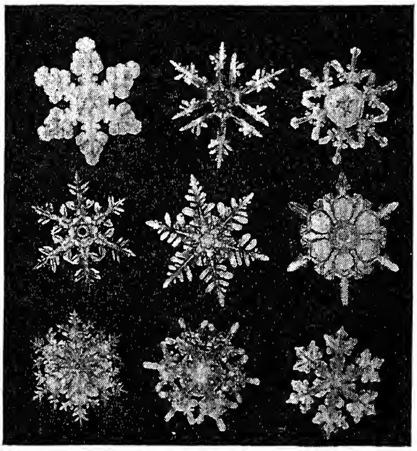

FIG. 160.—Snowflakes showing various patterns of snow crystals. (*Courtesy American Museum of Natural History.*)

Snow.—When the atmosphere becomes saturated with moisture, a slight drop in temperature produces condensation. If the temperature is above freezing, this excess moisture is eliminated in the form of rain; but when the temperature is less than 32°F., commonly it takes the form of snowflakes (Fig. 160). A large portion of the moisture that is precipitated as rain runs off as surface water, but snow generally accumulates in the region where it falls and generally remains as long as the temperature of the air in contact with it is below the freezing point.

Snow Fields.—In regions where the mean annual temperature is near the freezing point of water, much of the snow remains unmelted from one year to the next and consequently accumulates to great depths. The regions where such low temperatures commonly prevail are chiefly in high latitudes or at high elevations (Fig. 161). However, the amount of precipitation or snowfall is an additional factor and, apart from temperature, is perhaps the most important.

Where the amount of snow that accumulates during the cold season is greater than the amount removed by melting during the warmer season, a snow field is formed. Snow fields are widely distributed.

They may occur in any latitude at high altitudes and at all altitudes in high latitudes. Snow fields are common in the high mountains of South America and Mexico and in the Rocky Mountain system of the United States, Canada, and Alaska. They become progressively more wide-

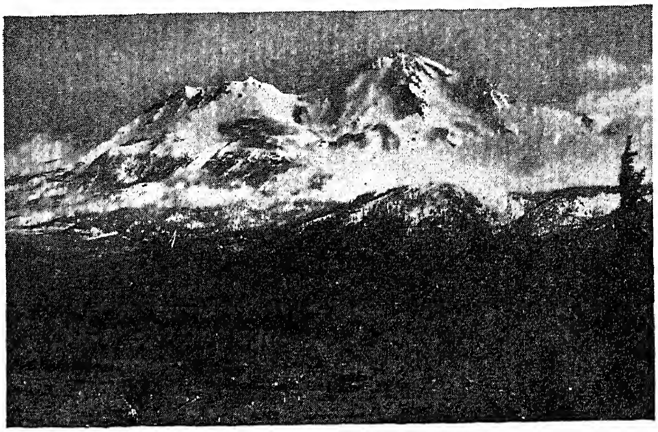

Fig. 161.—Mount Shasta, California, showing the snow fields and valley glaciers on its slopes. (*Courtesy Southern Pacific R. R.*)

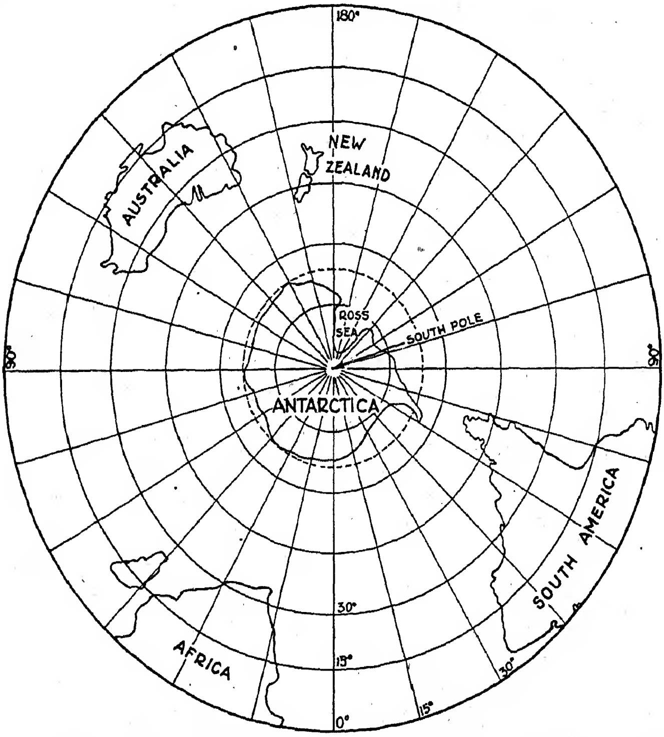

Fig. 162.—Map of the south-polar region showing the outlines of the continent of Antarctica and its size as compared with other continental masses. An ice sheet covers most of the polar continent.

spread toward the north. In Eurasia vast snow fields exist in the Himalayas, the Caucasus, the Pyrenees, and the Alps, and they also exist in the ranges of the Scandinavian peninsula. The two largest areas of snow and ice on the surface of the earth are those of Greenland and Antarctica, and of these Antarctica is by far the larger and contains more ice and snow than all the glaciers and snow fields of the rest of the world combined (Figs. 162, 163, 164). It has been estimated that more

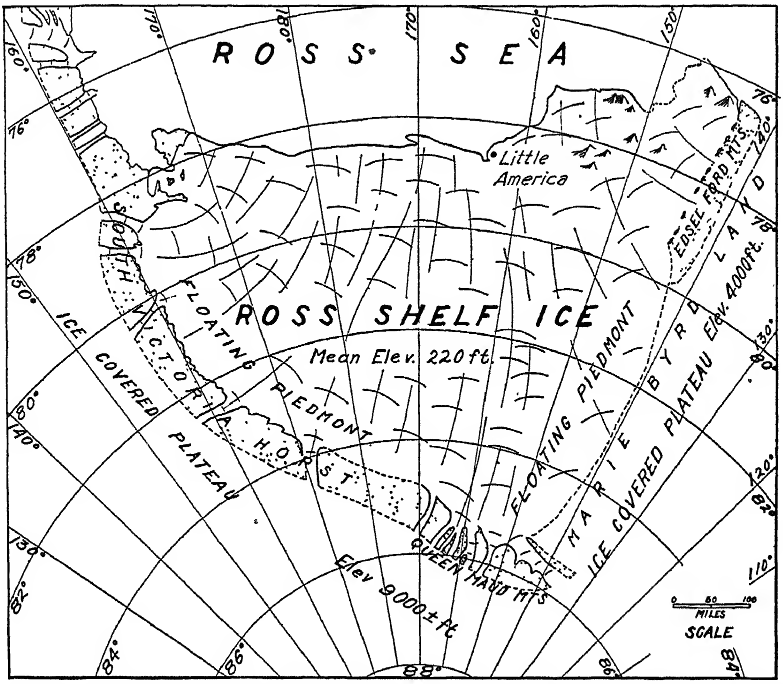

FIG. 163.—Map of the Ross Shelf Ice, Antarctica (see Fig. 162). A floating mass of ice approximately a thousand feet thick and covering more than a quarter of a million square miles. (*Redrawn from a map by L. M. Gould.*)

APPROXIMATE POSITION OF SNOW LINE WITH RESPECT TO ELEVATIONS ABOVE SEA LEVEL

	Feet
Greenland	2,200
Lapland	3,000
Norway	5,000
Pyrenees	6,000
Alps	9,000
Himalaya Mountains, south side	13,000
North side	16,000
Mexico	14,000
Andes, in Bolivia, east side	16,000
West side	18,000

than 5,000,000 cubic miles of snow and ice exist on the land areas of the earth at the present time and that if it were all melted it would raise the level of the sea approximately 150 feet.

Compacting of Snow.—The snow as it falls on the snow fields is commonly flakey and dry, but as it accumulates it is compacted, partially melted, and in refreezing acquires a granular texture much like small pellets of hail. Such granular snow is spoken of as *névé,* or *firn.* The transformation from snow to névé and this, in turn, into ice is very slow and is brought about by the interaction of a number of factors that result in the larger granules growing at the expense of the smaller ones. The loose, feathery snow gradually develops a texture similar to that of coarse sand or gravel. Compression and crystal growth eventually interlock the névé grains, forming a solid mass of ice. The process may be compared with what takes place when one makes a hard snowball. When the snow is squeezed into a ball, some of the little crystals of ice of which the snow is made are melted, and the ball becomes wet. As soon as squeezing is stopped, the water freezes and cements together the tiny crystals of snow. In the snow fields the weight of the upper layers of snow, added by successive storms and drifts or by avalanches

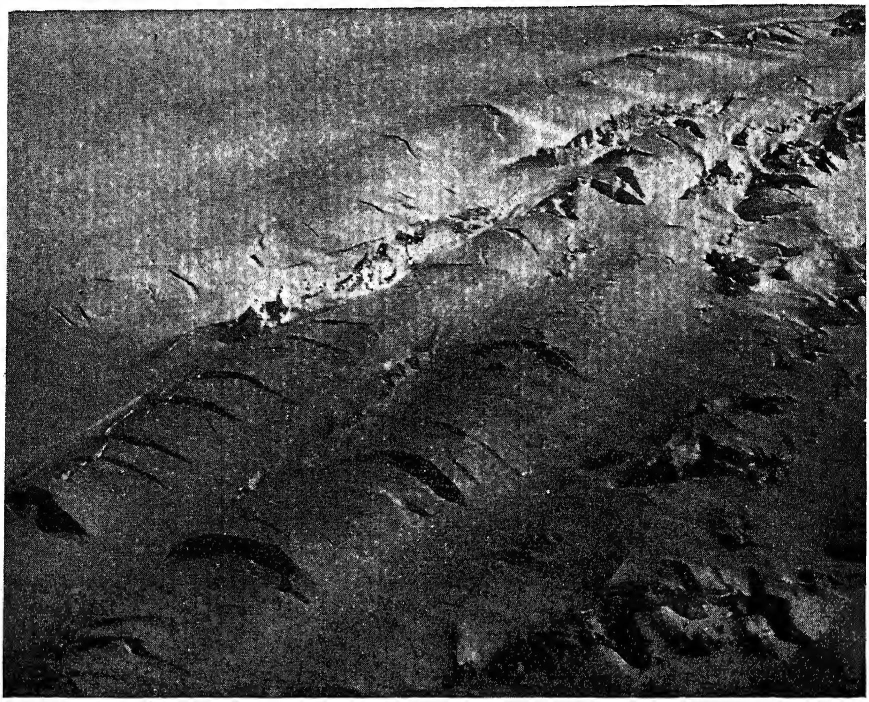

FIG. 164.—Folds in the Ross Shelf Ice over the Bay of Whales near Little America, Antarctica. (*Photograph courtesy L. M. Gould, Geologist, Byrd Antarctic Expedition.*)

from neighboring peaks, squeezes the lower layers, and they are recrystallized into solid ice.

During the summer months the surface of a snow field frequently reaches a temperature sufficiently warm to melt most of the flakey snow above the névé bed, and a thin crust of ice is formed when the temperature is again lowered. This crust separates the névé below from the snow of the next winter and remains in the névé field as a thin layer of ice; such layers give the whole deposit a stratified structure. Stratification is still more marked where wind-blown dust and coarser débris accumulate on the surface and are covered by subsequent snow-falls. In some Himalayan snow fields the separate strata of névé between successive layers or crusts of ice indicate an enormous amount of precipitation annually. In others, such as the firn fields of the Alps, each yearly deposit forms a thin layer from 2 to 5 feet thick. In some Alpine snow fields the total thickness of the névé beds is over 1,500 feet.

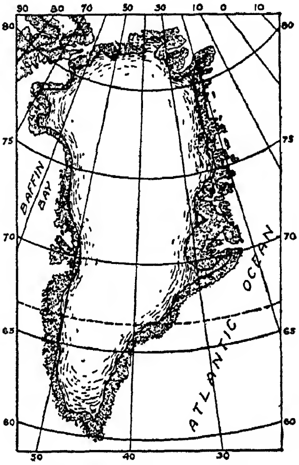

FIG. 165.—The continental ice sheet on the surface of Greenland. The main part of the island is covered by glacial ice, and there is only a narrow fringe of land along the coast.

Glaciers.—When granular snow or the massive ice resulting from its recrystallization accumulates to a great thickness, it begins to move or flow from its place of accumulation and forms a glacier. Such ice masses in mountain ranges vary in size from short but wide tongues of ice or snow on the narrow benches of a cliff, such as the snow-bank glaciers of the Bighorn Mountains of Wyoming, to the Seward Glacier of Alaska which is more than 50 miles long and over 3 miles wide. The thickness of the ice can only be estimated, but from the configuration of the surface it is safe to say that the ice in some of the mountain glaciers attains a thickness of nearly 2,000 feet. Still greater glaciers are found in the polar regions, where extensive névé fields accumulate, such as that of Greenland, which covers an area of nearly a half million square miles (Fig. 165). A part of Iceland also is covered by glacial ice (Fig. 166).

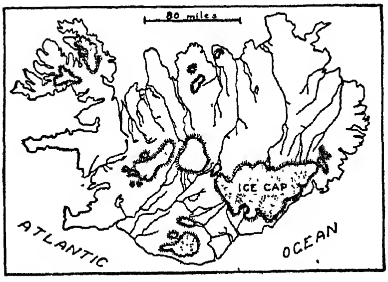

FIG. 166.—Ice caps on the surface of Iceland.

Types of Glaciers.—As glacial ice spreads or flows from the place where it formed, it assumes various shapes that are molded to a considerable extent by the surface over which the ice flows. On the basis of their mode of occurrence all glaciers may be classed as follows: (1) the mountain, or valley, type; (2) the confluent, or piedmont, type; and (3) the continental, or ice-sheet type.

Mountain glaciers skirt high peaks (Fig. 167) and occur along the flanks of mountain ranges in nearly all latitudes. They represent the discharge or solid drainage from an ice field through valleys which were originally formed by streams. At many places it is difficult to draw a definite distinction between a true glacier and a snow field, since they merge into each other, and one mountain glacier is formed by the union of a number of smaller tributary ice tongues each of which is in turn fed by the snow field.

Fig. 167.—Valley glaciers in the Selkirk Mountains in British Columbia. (*Courtesy Can. Geol. Survey.*)

If a mountain glacier descends to the lower end of a valley, it expands and spreads laterally as a broad, flat lobe or ice apron. Many Alaskan ice streams show such expansion; and when several glaciers from the same range coalesce in their expanded portions, a piedmont glacier is formed. The Malaspina Glacier in Alaska, a magnificent example of this type, is nourished by many ice streams from near the summit of Mount St. Elias and neighboring mountains. Continental glaciers or ice sheets cover vast areas and move over the lower ridges, hills, and valleys with but little regard for them. Most of Antarctica is now buried under such an ice sheet, and in the interior all the irregularities save the mountain peaks are concealed.

VALLEY GLACIERS

Distribution.—The most typical mountain glaciers are formed in well-defined valleys upon the mountain sides, and for this reason they are referred to commonly as valley glaciers. Their form is, in general, similar to the shape of the valley, and their size is dependent upon the extent of the snow field, the amount of precipitation, and the temperature.

Mountain glaciers are not present in all mountains which rise above the snow line, for on many isolated peaks the topography is unfavorable to the accumulation of snow. Extinct or dormant volcanoes may have great masses of snow and ice within their craters but few, if any, Alpine glaciers on their slopes. However, many ancient volcanic cones that were deeply dissected by streams before the last glacial epoch are now harboring hundreds of ice streams in their steeply dipping valleys. Such lofty peaks as Mount Shasta (Fig. 168), Mount Rainier, and Mount Hood send down large glaciers in all directions. On Mount Shasta some are 2 miles long, and on Mount Rainier several glaciers are nearly 7 miles long.

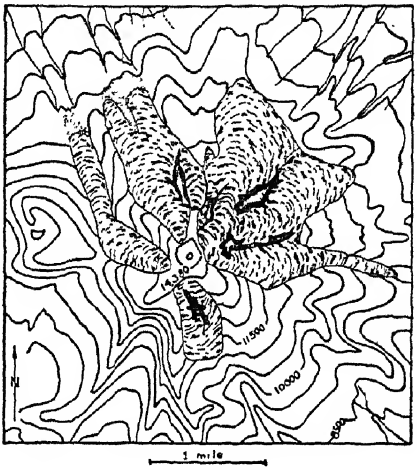

FIG. 168.—Glaciers radiating from the peak of Mount Shasta, showing the greatest number of glaciers and the lower elevation of the ends of the glaciers on the north side of the mountain. (*U. S. Geol. Survey.*)

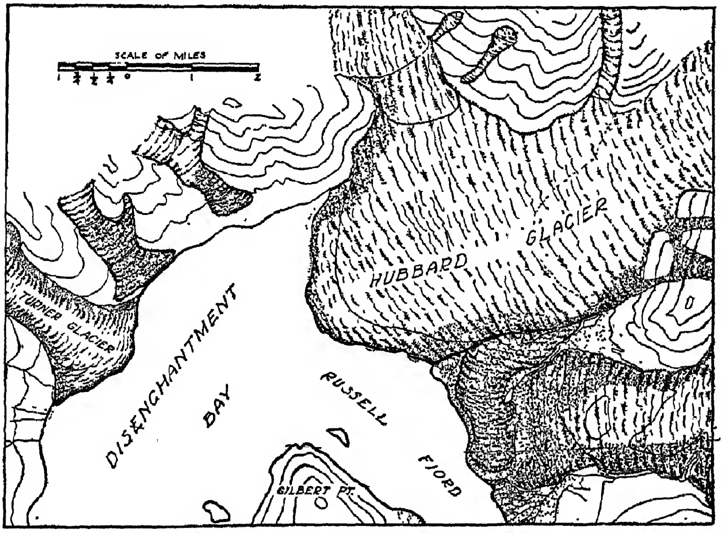

FIG. 169.—A group of glaciers in the region of Yakutat Bay, Alaska. These glaciers were once tributaries of a larger ice stream that filled Disenchantment Bay. (*After Tarr and Martin, Nat. Geog. Soc.*)

Valley glaciers occur in most of the high mountain ranges of the world. The Alps alone have approximately 2,000 in the depressions near the tops of their lofty peaks. In northern Scandinavia several large plateaus have glaciers moving into their marginal valleys. The Himalayas in Asia are famous for their huge ice streams, some of which are 30 miles long. In the United States the valley glaciers are confined to the isolated mountain peaks, but in the Selkirks of British Columbia and farther north in Alaska (Figs. 169 to 171) hundreds of glaciers of this type fill mountain valleys.

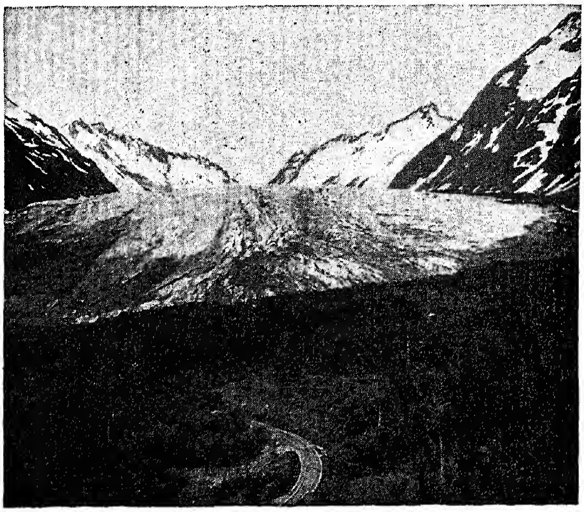

Fig. 170.—Terminus of the Bartlett glacier in Alaska. (*Courtesy Alaskan Railway Co.*)

Limits of Glaciers.—The position of the upper end or head of a valley glacier in general is determined by the gradational work of the glacier itself. As the glacial ice moves away from the snow field it carries with it large masses of rock plucked from the mountain side, and as this process is repeated a broad depression with the outline of an amphitheater is developed. Such an amphitheater, or *cirque*, commonly has precipitous walls hundreds of feet high, and its basin serves as the collecting ground for the snow of successive storms and for that swept into it by winds or carried into it by avalanches from the snow fields above. The

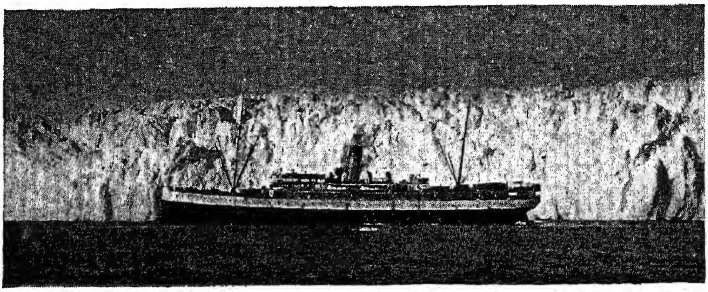

Fig. 171.—The terminus of the Columbia glacier, a tidewater glacier near Valdez, Alaska. The jagged ice cliff towers high above the ship's funnel. (*Photograph by H. W. Steward.*)

cirque wall, therefore, becomes the upper limit of the glacier, and its basin serves as the feeding ground for its ice.

The lower limit, or terminus, of the valley glacier generally is in the mountain valley, where the amount of ice waste or melting is about equal to the forward movement or flow of the ice. In most latitudes this position is some distance below the general snow line of the area; but as one proceeds toward high latitudes, the glaciers extend progressively to lower and lower altitudes, until near the polar regions they push downward to the sea where large blocks are broken off and floated as icebergs (Figs. 171, 172). The seaward end of a glaciated valley that is partly submerged is a *fjord* (Fig. 173). Fjords are typically developed along the coast of Norway. The end of a glacier is rarely stationary. When

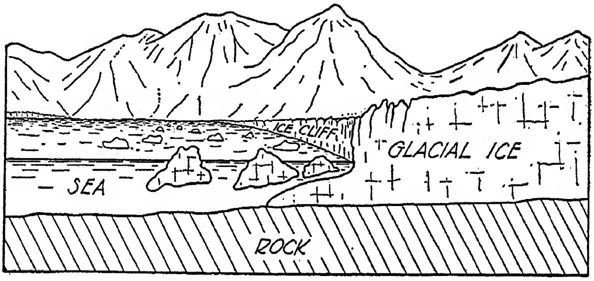

FIG. 172.—The end of a tidewater glacier. Huge blocks of ice break from the ice cliff and float away as icebergs. (*After Russell.*)

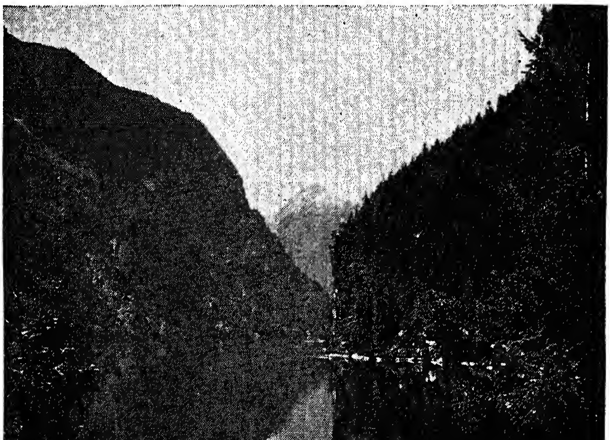

FIG. 173.—Fjord produced by glacial action on the coast of Alaska. (*Photograph by Curtis.*)

the temperature becomes such that melting exceeds the forward movement, the edge of the glacier retreats; and when more ice moves down than is melted, the edge of the glacier advances. The ice moves forward at all times, but the position of the edge depends on the rate of advance and the rate of melting. The amount of advance or retreat is dependent upon temperature and snowfall; and since both usually show annual variations, the lower ends of glaciers fluctuate. The most marked variations seem to appear with climatic cycles. The Swiss glaciers showed a steady advance during the Middle Ages and reached a maximum about 1820. This advance was followed by a progressive retreat until about 1840, when they again advanced until 1860, and since that time many have shown a marked retreat. However, during the past few years the glaciers of the French Alps have been advancing from 70 to 150 feet per year. Monthly observations on the Bossons glacier on Mont Blanc show that this glacier advances with an oscillating movement—first on one side, then on the other, then in the middle. A recession of the edges of glaciers in Glacier Bay, Alaska, is now in progress on a large scale, and it has been estimated that the edges of the Muir glacier have retreated 7 miles in the past 20 years (Figs. 174, 175).

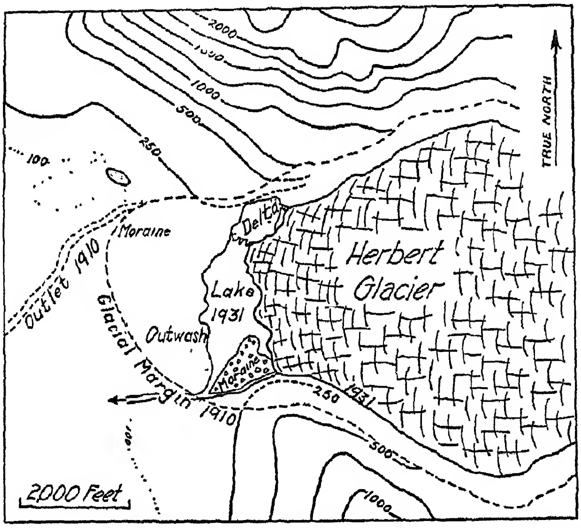

FIG. 174.—Map of Herbert glacier, Alaska showing the position of the ice front in 1931 as compared to the position in 1910. The 1910 outlet channel is abandoned. (*After C. K. Wentworth.*)

Movements of Glaciers.—A glacier moves so slowly that to a casual observer it appears to be at rest. By sighting at different times on rocks or other objects on the top of the glacier it is seen that these are moving slowly with the ice toward the lower end of the glacier. The exact nature of such movement is not yet thoroughly understood. The first determinations as to the amount of movement were made by noting the changes that took place in the débris on their surfaces. Conspicuous boulders on the surface of the ice were observed to change their positions slowly from year to year. Later such crude observations led to careful

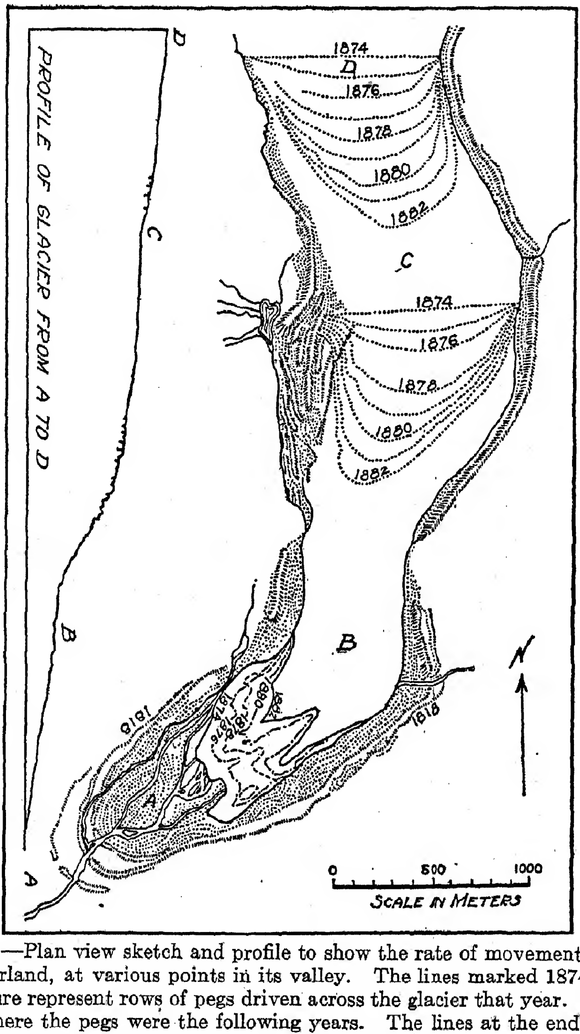

FIG. 175.—Plan view sketch and profile to show the rate of movement of the Rhone glacier, Switzerland, at various points in its valley. The lines marked 1874 in the upper part of the figure represent rows of pegs driven across the glacier that year. The numbers below show where the pegs were the following years. The lines at the end of the glacier show the edge of the ice in the year stated. (*After Heim.*)

measurements of the rate of change, which was found to vary considerably not only in different glaciers but in different parts of the same glacier.

The rapidity of glacial flowage is influenced by several factors, the chief of which are the gradient of the valley, the thickness of the ice, and the temperature. To these may be added the smoothness of the surface over which the ice moves, the amount of water in the ice—both that which falls on it and that produced by melting—and finally the amount of débris in the ice. If a glacier is loaded with boulders, sand, and finely ground rock waste, it moves more slowly than clean ice.

Alpine glaciers move from 1 to 3 feet a day. The Bossons glacier carried the bodies of three guides, who perished in a fissure, a distance of nearly 8,000 feet in 41 years, or an average of about 1 foot in 2 days. It has been calculated that a particle of ice would require approximately 500 years to move from the summit of the Jungfrau to the end of the Aletsch glacier, an ice stream about 10 miles long. Many of the large Alaskan glaciers move at a more rapid rate. The Muir glacier commonly moves as much as 7 feet a day, and the Child glacier flows nearly 30 feet per day during the summer months. Extraordinary velocities have been recorded for the ice tongues which descend to the fjords along the coast of Greenland, where rates of nearly 100 feet a day have been observed, but in the same region the inland ice at some distance back from the narrow fjords moves only a fraction of an inch per day.

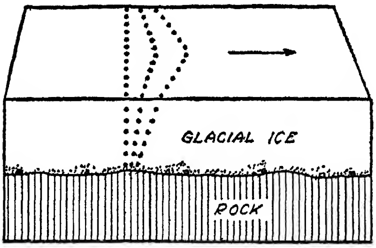

Fig. 176.—Diagram illustrating the differential movement of a valley glacier. Stakes driven in a straight row across the top of a glacier, in a short time show a curve downstream. Pegs placed on the side of the glacier demonstrate that the top moves faster than the bottom.

The movement of a glacier resembles that of a river in many ways; the center moves more rapidly than the margins, where it meets resistance along the walls of the valley; the surface moves more rapidly than the deeper portions of the ice, where its load of débris is greater and irregularities are encountered on the valley floor. At curves in its course the convex portion moves more rapidly than the concave, in much the same way that the velocity of a stream is accelerated on a long limb of a curve and checked on the opposite side. Since some parts of the ice move faster than other parts, the movement of a glacier is spoken of as "differential movement" (Fig. 176). The amount of differential movement has been determined in many glaciers by placing stakes in a straight line over the surface of the ice in alignment with fixed points on the walls of the valley. After a few days the line of pegs curves downstream, indicating that the central part moves more rapidly than the sides.

At many places near the center of a glacier the ice moves four times as fast as it does near the sides. Similarly, by driving a vertical line of pegs where a wall of ice is exposed at the side of a glacier, it is found that the top moves faster than the bottom (Fig. 176).

Several theories have been proposed to explain glacial movement, and the problem has been discussed with the aid of extensive experimental data, but it is beyond the scope of this text to review the data here.

Formerly it was thought that a glacier flows like a very viscous fluid that spreads out under its own weight. Later work, however, indicates that there are other factors, one of which may be the melting of the ice at points of greatest compression and the refreezing of the water produced. When it refreezes, it expands, and the surrounding ice mass is subjected to the thrust of its expansive force. It is evident that the thrust in the direction toward the lower end of a valley is augmented by the force of gravity, whereas movement toward the head of the glacier is retarded by the same force. Furthermore, as glacial ice moves, it recrystallizes and develops a structure somewhat similar to that of a schist such as is produced by the recrystallization of a shale. The ice crystals become arranged in a more definite order with their axes tending to become parallel. As this process continues, gliding planes are developed in the crystals and along the crystal faces, and movement takes place along such planes. Where the lower portion of a glacier carries the most débris, it is greatly retarded by its load, and the clean ice above moves so much faster that it is sheared over the portion near the bottom. Such shearing planes are conspicuous in the glaciers of high latitudes, where vertical or overhanging ice cliffs hundreds of feet high allow the structure of the ice to be seen.

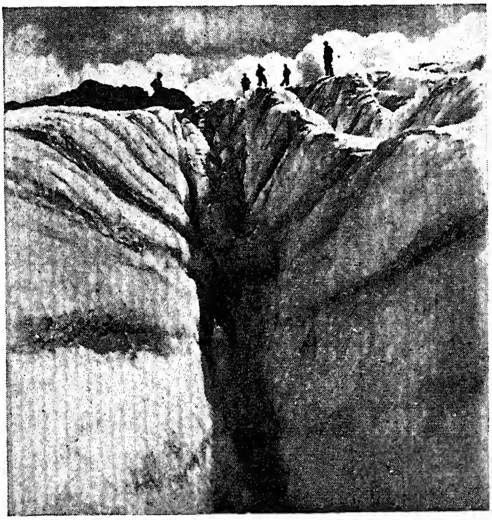

FIG. 177.—Crevasses on the surface of Sperry glacier, Glacier National Park, Montana. (*Courtesy Great Northern R. R.*)

Surface Features. *Crevasses.*—Where glacial ice passes over irregularities in the bottom of the valley, a change from a lesser to a greater gradient is encountered, and tension is produced in the upper surface so that the rigid ice cracks. The resulting fissures are called *crevasses*

(Fig. 177). Their direction on the surface of the glacier generally is roughly transverse to the long axis of the ice stream. When first formed, they extend downward nearly vertically; but since the upper surface of the glacier moves more rapidly than the lower portion, they assume an inclination which dips up the valley. They also curve downstream because of the more rapid movement of the ice in the center of the valley. Such fissures are widest in the central portion of the glacier and taper gradually to narrow cracks at their extremities. They vary in depth from a few to several hundred feet and in width from a fraction of an inch to great chasms. In some it is safe to descend to the bottom, and in those near the upper end of the glacier where englacial débris is not abundant the wall ice is perfectly clear with a greenish-blue transparency. The sides of the crevasses are frequently hung with icicles and embossed with wreaths of snow.

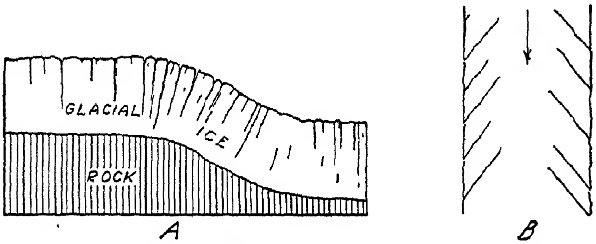

FIG. 178.—Diagrams illustrating the origin of crevasses. *A*, crevasses produced by cracking due to a change in the gradient of the bed of the glacier; *B*, marginal oblique crevasses produced by the more rapid movement of the central part of the glacier, which tends to pull the ice from the more slowly moving marginal portions.

Along with the transverse crevasses most glaciers are fissured also on the sides or margins (Fig. 178). The marginal crevasses do not extend in the direction of the long axis of the glacier but point up the valley about 45 degrees. They are due to the more rapid movement of the central portion of the glacier. Longitudinal crevasses occur also, especially wherever a glacier issues from a narrow portion of a valley into a wider one. There it has room to spread and in so doing has a tendency to fall apart, forming longitudinal fissures. Similar structures are common also on terminal lobes.

Bergschrund.—At the upper end of a mountain glacier where the ice breaks away from the névé fields, there is a great crevasse or a series of open fissures that is known as the bergschrund (Fig. 179). The névé and ice forming the upper margin of the bergschrund stand higher than the portion that has moved away. The displaced block has been subjected to a downward and to a horizontal movement as well, thus forming a huge crack in the ice. Such movement is greatest during the summer months. During the winter the process is halted, and the bergschrund fills with snow and ice that enters the irregularities and joints of the rock wall. The following spring it opens again and carries big blocks of rocks with it. In this way huge amphitheaters are formed at the upper ends of glacial valleys.

Moraines.—The rock débris carried by glacial ice is morainic material, and from its position with respect to the ice mass it is classified as superglacial, englacial, or subglacial morainic débris. Its distribution on and in a glacier is dependent upon the topography of the mountain valley. If the valley is deep, the mountain sides are undermined by the moving ice, and heaps of rock fragments or immense blocks of rock gather on the ice at the sides of the valley, forming a lateral moraine; and when two tributary glaciers meet, a medial moraine (Fig. 180) is formed by the union of the lateral moraines. As a rule the medial moraine loses its identity toward the terminus of the glacier. Often, however, the moraines remain distinct and may be seen for miles, stretching up the glacier side by side.

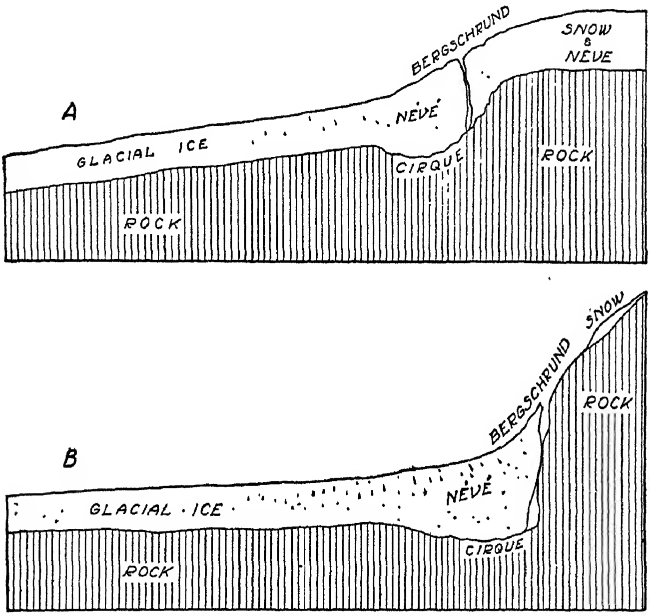

FIG. 179.—*A*, cross section of the upper part of a valley glacier showing the development of a cirque. The bergschrund lies between the head of the glacier and the snow field. It consists of one or more crevasses from 1 to 50 feet wide. *B*, glacial plucking has excavated a deep cirque with a steep rock wall which serves to catch snow and névé that feed the glacier.

The débris carried to the terminus of the glacier is deposited about its end as the ice melts and at many places forms a crescent-shaped ridge known as a terminal moraine. Such ridges are more pronounced in continental and piedmont glaciers where the glacier-fed streams have lower gradients and, therefore, do not so readily carry away the glacial débris. Mountain streams have high velocities, and they therefore carry the material of the terminal moraine for some distance beyond the glacier and distribute it along the valleys as they aggrade their beds.

Huge heaps of ice blocks often are found where a tributary glacier unites with the main ice stream by descending a precipice. Large masses of ice are detached at the end of the tributary and are dropped to the glacier below, forming *ice falls* and *cascades.* Later the shattered blocks are recemented at the base of the cliff and become incorporated into the mass of the main glacier.

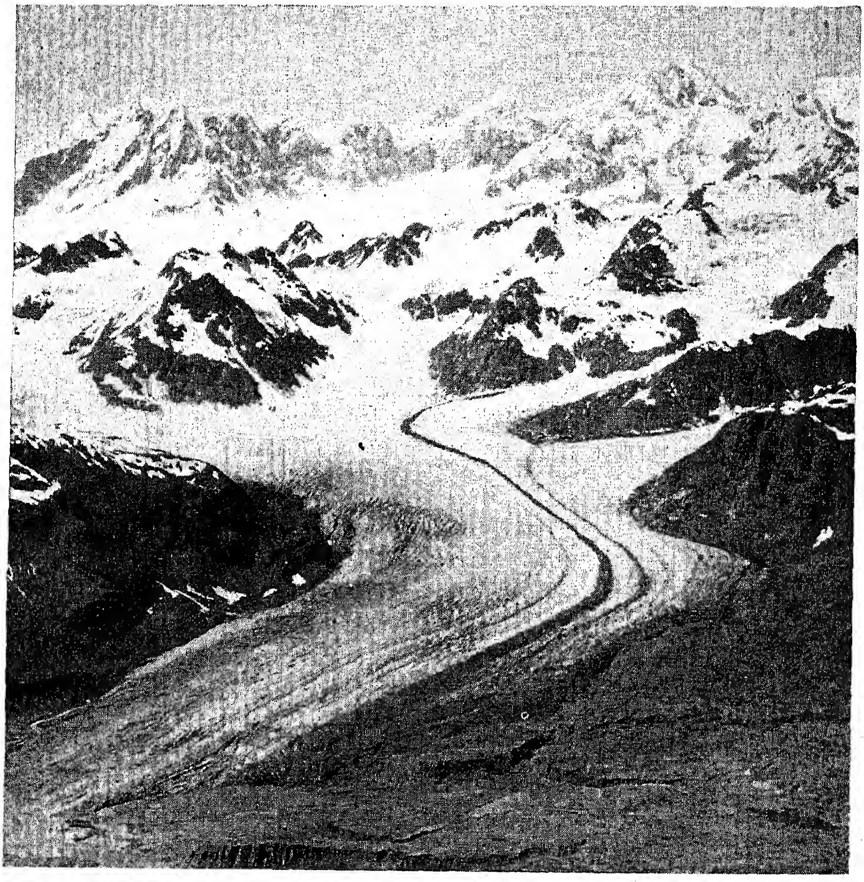

Fig. 180.—Geikie glacier, Alaska as seen from the air. Medial moraines derived from the lateral moraines of the tributary glaciers to the right, appear as dark lines in the ice. (*Courtesy U. S. Forestry Service.*)

Features Due to Ablation.—The surface of a glacier is exposed to the heat of the sun and the action of dry winds, both of which produce irregularities by melting and evaporation. Morainic material on the glacier protects the ice from the sun's rays so that at places the moraines or belts of débris lie on ice walls that project upward 100 feet above the general surface of the ice. Such ridges are especially conspicuous on the Aar glacier in Switzerland, and in Greenland some have been found that are nearly 400 feet high. In the same way, flat blocks of rock shelter the ice beneath and remain on pillars or pedestals as the surrounding surface is lowered. Such structures are spoken of as "ice tables," or "glacier tables" (Fig. 181). The tables are rarely horizontal but are inclined to the side that is most exposed to the rays of the sun. As melting continues, the blocks of stone perched on the columns of ice are dislodged, and the pinnacles that formerly supported them remain on the surface of the glacier as irregularities called "ice pyramids."

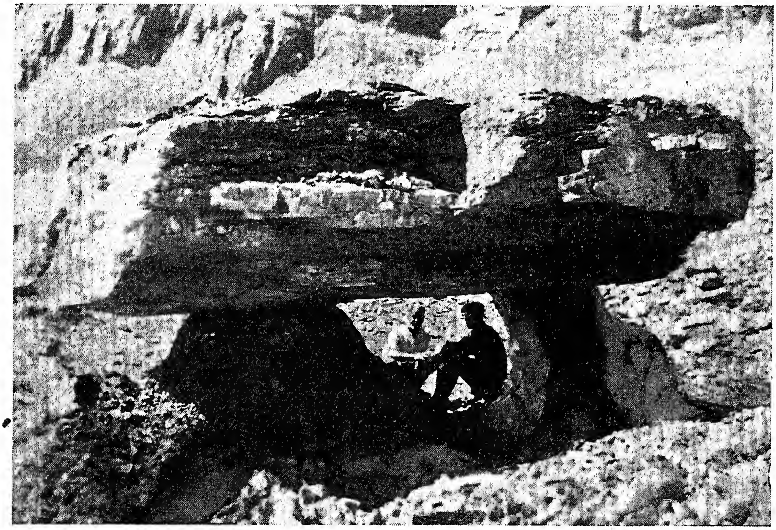

FIG. 181.—Ice table at the foot of Grinnel glacier, Glacier National Park, Montana.

Small heaps of dust and thin slabs of rock absorb the sun's heat and so become sufficiently warm to melt the underlying ice. They sink below its surface, forming depressions, or *dust wells*. For the same reason thin, bouldery moraines sometimes are found sunk below the surface of the glacier, or scattered pebbles are found at the bottom of water-filled pits. Leaves are blown on the surface of some mountain glaciers, and during the summer months they are warmed by the sun and sink below the surface.

On warm days the surface of a glacier has innumerable rills and pools of water, which unite to form small streams that rush and tumble as waterfalls into the crevasses and are lost in the depths of the ice. The water carries with it mud, sand, and boulders from the surface moraines. These erode the crevasses and produce vertical shafts that are referred to as *glacier mills*, or *moulins*. A moulin once formed moves down the valley with the glacier, but since the irregularity in the floor of the valley that produced the crevasse in which the moulin formed remains stationary, the process is repeated many times near the same place, and a long row of deep shafts is developed in the ice. Some of these streams erode subglacial channels that form large tunnels in the basal part of the ice (Fig. 182).

Erosion by Valley Glaciers.—The results of erosion by valley glaciers are so characteristic and differ so widely from those produced by other gradational agents that a glaciated valley is readily recognized even though the glacier itself has entirely disappeared. The erosion may be accomplished (1) by cleaning off the residual, loose débris; (2) by breaking or wearing off the surface of the bedrock over which it passes; and (3) by a process known as plucking, whereby joint blocks are pulled out and carried along with the ice. Such erosion is accomplished not merely by the pressure of the ice but by means of the sharp sand, angular pebbles, and boulders and other rock débris that serve as the abrasives with which the glacier grinds and polishes. The intensity of this action can perhaps be appreciated more fully by considering the force exerted when a thick mass of ice passes over a rock surface. Since a cubic foot of ice weighs about 57 pounds, a glacier 1,000 feet thick would exert a pressure of approximately 28 tons per square foot.

FIG. 182.—Ice tunnel formed by subglacial stream. Terminus of Mendenhall glacier, Juneau, Alaska. (*Photograph by Griffith.*)

The boulders and pebbles left in a glaciated valley are polished and striated in such a manner as to indicate that ice is sufficiently rigid to hold the débris firmly. The embedded rocks act as the teeth on a gigantic file. Many glaciated valleys appear as if such a file had been pressed down heavily and dragged along the valley. The irregularities on the floor and walls of the valley are rounded, grooved, and smoothed on the upstream side and show sharp angular projections on the leeward side where plucking rather than abrasion was most active. Much of the waste of such glacial abrasion is carried away in suspension as finely ground rock flour that gives a milky appearance to the streams that issue from the end of a glacier.

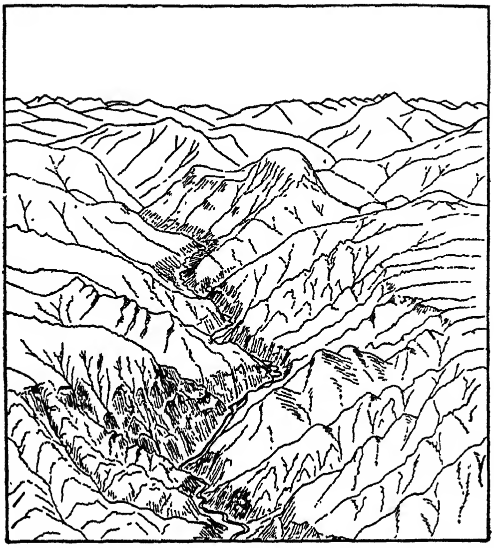

Fig. 183*A*.—Yosemite Valley as it probably appeared *before* it was glaciated. (*After Matthes, U. S. Geol. Survey.*)

Fig. 183*B*.—Sketch of the Yosemite Valley in California, after glaciation, showing a deep, flat-bottomed valley carved out of massive rock. The deepening accomplished by the ice ranges from 600 to 1,500 feet. A lake 5½ miles long once occupied the basin gouged into the rock floor of the valley. (*After Matthes, U. S. Geol. Survey.*)

Changes Produced in Topography.—Where glacial erosion is active for an extended period, ice-sculptured topographic features replace those

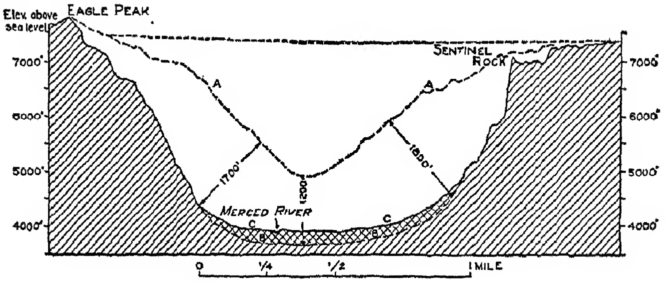

Fig. 184.—Cross section of Yosemite Valley from Eagle Peak to Sentinel Rock showing the deepening and widening accomplished by glacial erosion. *A-A*, probable preglacial profile; *B-B*, bottom curve of the glacial U-shaped valley; *C-C*, present profile. Depth of glacial excavation, 1,200 feet. (*After Matthes, U. S. Geol. Survey.*)

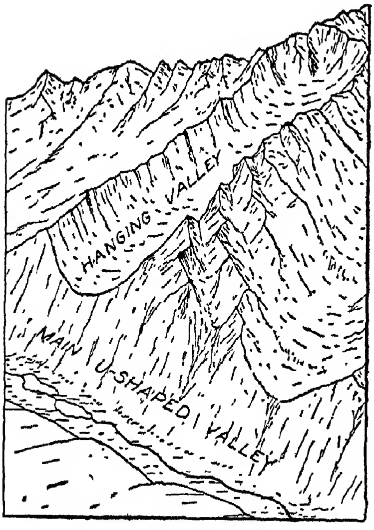

Fig. 185.

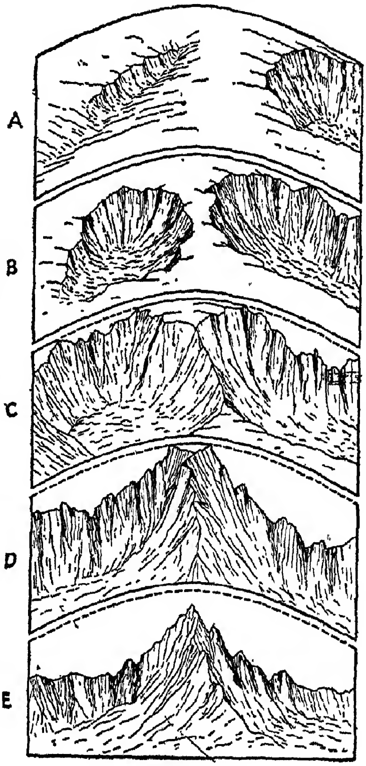

Fig. 186.

Fig. 185.—Diagram showing the relation of a *hanging valley* to the main valley.

Fig. 186.—Successive stages in the development of cirques. The headward growth of cirques narrows the divides separating them, until eventually sharp and jagged comb ridges remain. The high sawtooth pinnacle shown in *E* is called a *horn* or a *matterhorn peak* after the famous Matterhorn in the Alps. (*Redrawn from Davis.*)

produced by ordinary atmospheric weathering and corrasion. A youthful river valley is usually V-shaped in section; but when such a valley is glaciated, its sides are eroded as rapidly as the bottom, and a broad, flat-bottom, U-shaped valley results (Figs. 183 to 187). The projecting and overlapping spurs that are characteristic of youthful topography are truncated, and the walls of the valley are made smooth and straight. In Alaska many of the glaciated valleys are straight, and they are referred to as "canals."

The floor of a glaciated valley often shows irregularities rounded into dome-shaped bosses that are sloping and smooth on the side from which the glacier moved but rough and jagged on the leeward side as a result of glacial plucking. Such rock structures are known as *roches moutonnées* (sheep-shaped rocks).

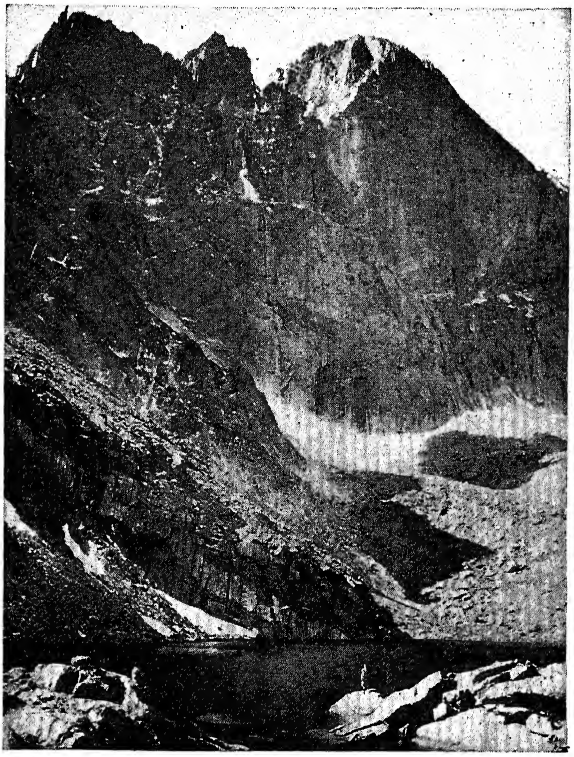

FIG. 187.—A cirque wall on the slopes of Longs Peak, Colorado. A towering granite cliff over 2,000 feet high, formed by glacial plucking. Chasm Lake occupies the cirque basin. (*Courtesy Denver Tourist Bureau.*)

In a normal river system the tributary streams enter the main valley at the level of the main stream. If such a system is glaciated, the main valley, because of its greater volume of ice, is deepened more rapidly than the smaller tributary, and when the ice disappears the elevation of the tributary at the junction is higher than that of the main channel. In many mountain streams the difference amounts to hundreds of feet, and in some of them to more than 2,000 feet. Valleys that discharge high above the floor of the main stream are called *hanging valleys* (Fig. 186), and their streams leap in a series of cascades into the major channel or discharge as waterfalls such as the famous waterfalls of the Yosemite.

Mountain glaciers develop *cirques*, or amphitheater-like depressions with steeply rising walls, at the upper ends of glaciated valleys (Fig. 187). These gigantic semicircular bowls carved out of the mountain sides are among the most striking features in the mountain topography of such regions as the Selkirks in Canada, the Rocky Mountains of the United States (Fig. 188), or the magnificent mountain scenery of the Alps in Switzerland. Their precipitous walls are produced by the plucking and sapping action of the ice, which adheres to the rocks by filling the joints and fractures near the base of the cliffs. During the warmer season, as the ice moves down the valley, it pulls away from the walls of the cirque and plucks out the blocks of rock so that they move with it. With the coming of winter the process is halted, and the crevasse, or bergschrund,

FIG. 188.—Aerial photograph of small cirques southwest of Berthoud Pass, Colorado. (*Photograph by T. S. Lovering, U. S. Geol. Survey.*)

between the moving ice and the rock wall fills with snow; but the following spring it again opens, and more rock is quarried from the cirque wall. In this way cirques are progressively enlarged, deepened, and caused to extend headward, until eventually those on one side of a mountain range recede into the area of others on the opposite side of the divide, and the space between them is narrowed, producing sharp, serrated, comb-like ridges (Fig. 189), or "horns," such as the lofty Matterhorn of the Swiss Alps. Most of the deep, clear lakes, that lend beauty to the mountains rest in rock basins that are parts of glacial cirques.

Features of Deposition.—Since mountain glaciers are surrounded by large areas of exposed rock that are constantly subjected to denudation, they collect a great deal of rock waste that is carried by the ice toward the terminus of the glacier. This débris, called moraine, is supplied chiefly by the mechanically weathered material which falls from the walls of the valley and by rock abraded from the bed, and it is classified with reference to its position as lateral or medial moraines or as superglacial, englacial (Figs. 190, 191), or subglacial débris. When the ice eventually is melted, both ice and water play a part in the deposition of the sediments. The heaps of débris dropped directly by the ice are called *glacial deposits*, and those made by glacial streams are called *glaciofluviatile* deposits. The term *glacial drift* is commonly applied to all of the material transported and deposited by glacial ice. *Till* is unstratified glacial drift.

FIG. 189.—Apache Peak, Colorado. A knife-edge ridge produced by the headward erosion in glacial cirques. (*Photograph by H. E. Kellogg.*)

Ground Moraine.—The basal portion of a valley glacier may become so filled with débris that the ice cannot transport it all, and then part of it remains upon the bottom and is overridden as the ice moves onward. It is most likely to become overloaded for some distance beyond a place where an irregularity at the lower surface of the ice favors the gathering of débris or near the end where the ice mass is thinner. All the material deposited beneath the advancing ice, together with that deposited from the base as an irregular sheet during melting, constitutes the ground

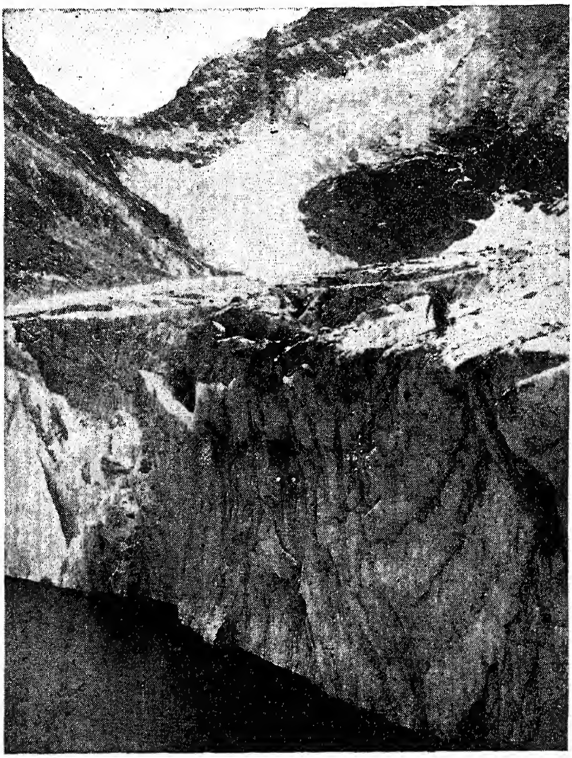

Fig. 190.—The terminus of Windermere glacier, British Columbia, where it discharges into a lake. Note the crumpled structure as outlined by the englacial débris. (*Courtesy Can. Department of the Interior.*)

Fig. 191.—Side view of Grasshopper glacier near Cooke, Montana. The ice contains millions of grasshoppers that have fallen in the snow and *névé* fields of the glacier. Note the flow lines and stratification produced by the differential movement of the ice. (*Photograph by T. S. Lovering.*)

moraine. It consists of a heterogeneous mass of fine clay or sand, striated pebbles, and boulders that show neither assortment nor stratification. The material of such ice-deposited wastes sometimes is called boulder clay, because it is made up of fine to coarse sandy clay through which are dispersed boulders of all sizes and up to many tons in weight.

Terminal Moraines.—At the terminus of a glacier where the amount of ice waste due to melting equals or nearly equals the advance due to glacial movement, débris is dropped as a *terminal moraine.* In many regions where the glaciers at one time pushed out upon the piedmont areas and halted there for a number of years, the moraines are still present as well-defined topographic features. They are usually crescent- or horseshoe-shaped in outline, with their concave sides toward the head of the valley (Fig. 192). The height of such a moraine is dependent to a considerable extent upon the length of time during which the front of the glacier remained stationary, for it is obvious that rapid retreat would not

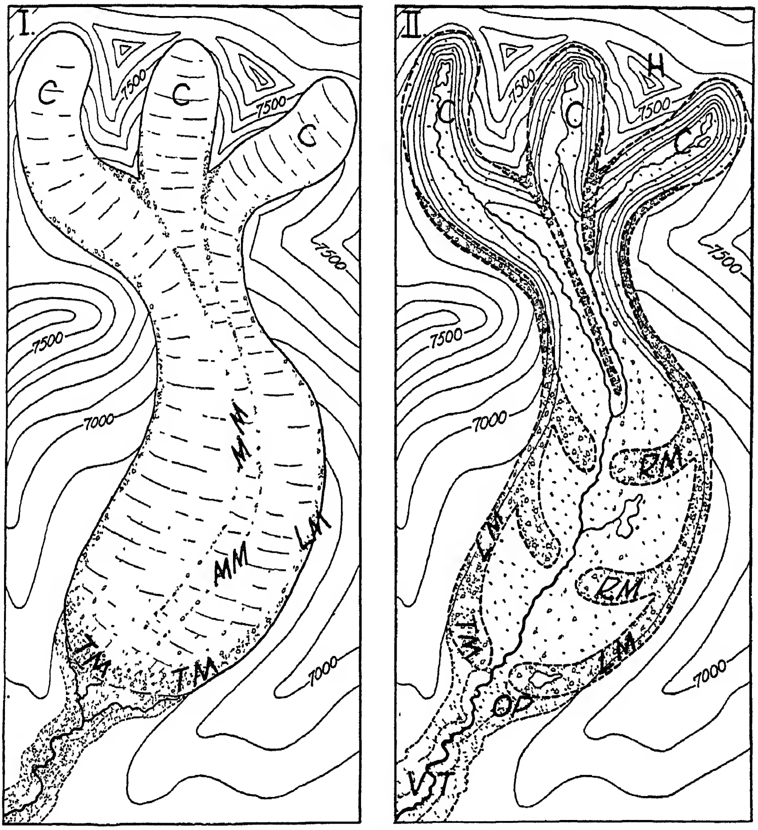

FIG. 192.—*I*, a mountain glacier; *II*, the same after melting of ice; *C*, cirque; *LM*, lateral moraine; *TM*, terminal moraine; *RM*, recessional moraine; *OP*, outwash plain; *VT*, valley train; *H*, horn.

allow much material to be piled up at the same place. If a glacier advanced 1,000 feet a year for a period of 50 years and melted as fast as the ice moved, the end would remain stationary, and all of its moraine débris would be deposited at approximately the same place in the valley during the whole period of time. However, if the end of the glacier receded 1,000 feet a year, even though its ice moved forward at the rate of 500 feet a year, very little débris would be piled up at any spot, and consequently no pronounced morainic ridge would be formed.

Lateral Moraines.—The débris that accumulates on the borders of a valley glacier forms the lateral moraines of the moving ice stream. When the glacier melts, the lateral moraines are left as ridges or terrace-like structures bordering the steep-sided mountain valleys (Fig. 192). In most glaciated valleys they are more conspicuous than either the ground moraines or the terminal moraines. Where the gradient of the mountain valley is not too great, the glacial streams leave most of the material of the terminal moraine at the place where it was deposited, and at such places the lateral and terminal moraines unite as a continuous ridge or dam across the valley, and the inclosed basin becomes occupied by a lake, swamp, or meadow.

Glaciofluviatile Deposits.—In the region where the melting of glacial ice is in progress, innumerable streams are formed on the surface, at the margins, and at the bottom of the glacier. During the summer months great torrents issue from ice caves or tunnels and carry boulders, pebbles, sand, and fine rock flour; but as the streams emerge from the restricted channels on or in the ice, they spread out over a greater area or divide into many distributaries and consequently become overloaded and drop a large part of their sediments. Since the streams issue from the ice in the region where the terminal moraine is being deposited, most of the glaciofluviatile sediments are carried beyond the terminal moraine. The shape of the water laid deposit where the glacier ends on a plain area is similar to that of an elongated, alluvial fan. Where the end of the glacier is confined by the walls of the valley, the fluviatile deposits are confined to the width of the valley and build a *valley train.* Its sediments are sorted and stratified with coarse gravel and boulders near the glacier, grading horizontally through fine sand to silt and rock flour at a distance of a few miles below the terminus of the glacier.

PIEDMONT GLACIERS

Piedmont glaciers are found only in subpolar regions in areas with mountains of strong relief, where, by the confluence of several glaciers, an ice plateau is formed upon the more nearly level country at the foot of the mountain range (see Fig. 193). Because of the marked change in gradient of the valleys as they reach the piedmont area, the movement of the ice is greatly retarded, and in many piedmont glaciers it becomes almost imperceptible. The Malispina glacier on the western side of Yakutat Bay in Alaska is a classic example of this type of glacier (Figs. 194, 195). It consists of three principal lobes, and each lobe represents

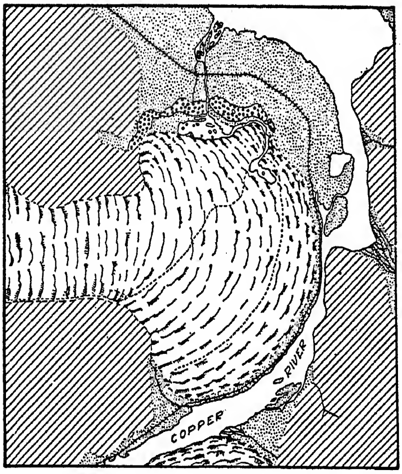

FIG. 193.—A bulb glacier showing the mode of origin of a piedmont glacier, Baird glacier, Copper River district, Alaska. (*Nat. Geog. Soc.*)

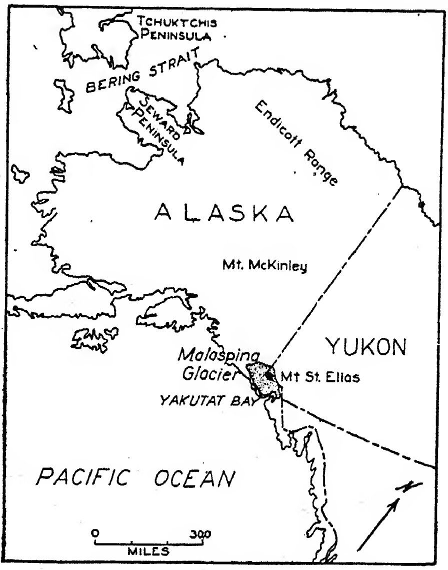

FIG. 194.—Map of Alaska showing the location of the Malaspina glacier, a typical piedmont glacier formed by a group of valley glaciers that descend from the slopes of Mount St. Elias.

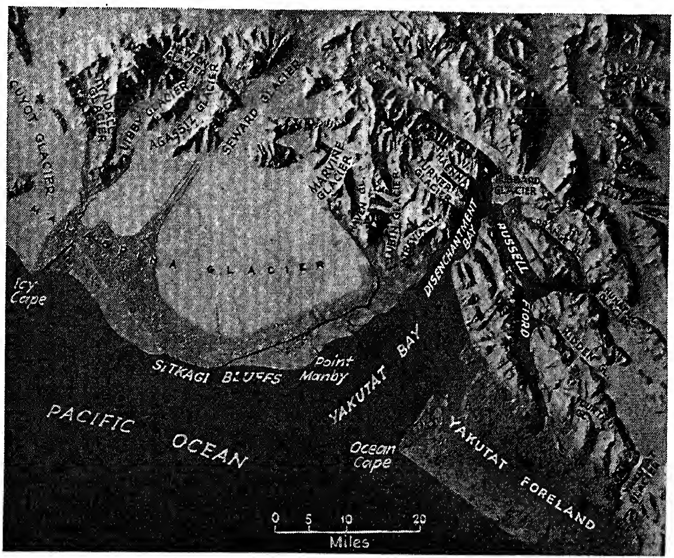

FIG. 195.—View of model of region including Yakutat Bay and Malaspina glacier. (*Model by Lawrence Martin. Copyright*, 1909, *by University of Wisconsin.*)

the expanded lower portion of one or more large valley glaciers that move down the slopes of Mount St. Elias. These lobes are fed by the Seward, Agassiz, Tyndall, and Guyot glaciers, and the piedmont glacier formed by their coalescence is a vast, nearly horizontal plateau of ice that is 70 miles wide along the Alaskan coast. Its extension inland is from 20 to 25 miles, and its area is approximately 1,500 square miles. Where the individual valley glaciers emerge from their mountain valleys, differential movement is initiated in the ice plateau, and its surface becomes broken by thousands of crevasses. Toward the margins, where ice movement decreases and melting and freezing increase, these crevasses are gradually healed, and the surface shows but minor irregularities. The outer margin for a width of 5 miles or more is covered with moraine. At certain places this marginal ice has reached such a stage of stagnation that forests grow upon the moraine, even though ice hundreds of feet thick is still present below the morainic soil.

CONTINENTAL GLACIERS

The valley glaciers are generally of restricted size, and as noted they are usually much longer than they are wide. Other glaciers, however, have great width, and some are of such great size that they approach continental extent and are called *continental glaciers*. They are great sheets of ice rather than ice streams. The Antarctic continental glacier has an area of more than 2,000,000 square miles. In the continental glaciers there are no visible feeding valley glaciers, but the whole ice sheet presents a featureless and very gently sloping surface, except here and there where a bare hill or peak known as a *nunatak* rises above the general level of the ice surface. The continental ice sheets are not fed by small tributaries with marginal belts of ice and rock. There are no medial moraines in the continental glaciers corresponding to the medial moraines of the valley glaciers, but above the zone of melting the ice is clear and white.

Erosion by Continental Glaciers.—The efficiency of a continental ice sheet as an instrument of erosion is particularly well shown by the effects of the vast ice cap that covered northeastern North America during the geologically recent ice age. From a study of the features developed during that period, it is evident that corrasion is by far the most important erosion process and that it is accomplished primarily by the scouring and grinding action of rock fragments either frozen into the bottom of the glacier or pushed along beneath it (Fig. 196). Glacial plucking is limited in its action to areas of steeply dipping or highly jointed rock formations.

A feature peculiar to extensive lowlands that have been glaciated by ice sheets is the almost total absence of residual soils or mantle-rock waste. Such loose soils and rotten rock are scoured off and dragged along by the passage of the ice, and bare, rounded, and striated surfaces of fresh rocks remain (Figs. 197, 198). Similar features are formed also by mountain glaciation. In some regions, like the plains of the upper Mississippi basin, the glaciers moved over large areas of residual soils with little disturbance of them, but such soils are now covered with a thick bed of glacial till.

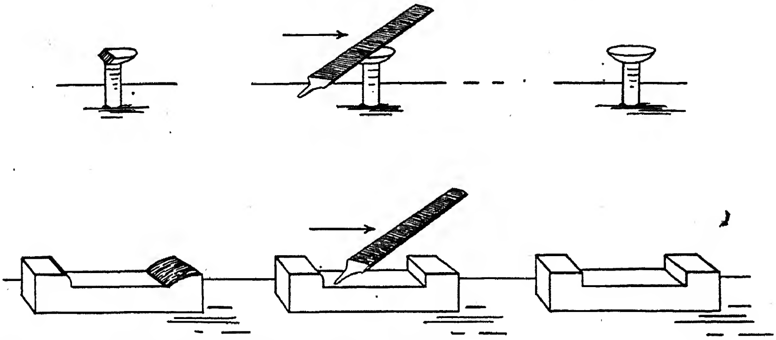

FIG. 196.—Diagram showing the effect of a file moving over a nailhead and over a depression. The abraded side of the nail is toward the direction from which the abrading agent came. In the depression, as shown by the blocks, the part most abraded is in the direction toward which the agent moved. The figure illustrates ice erosion.

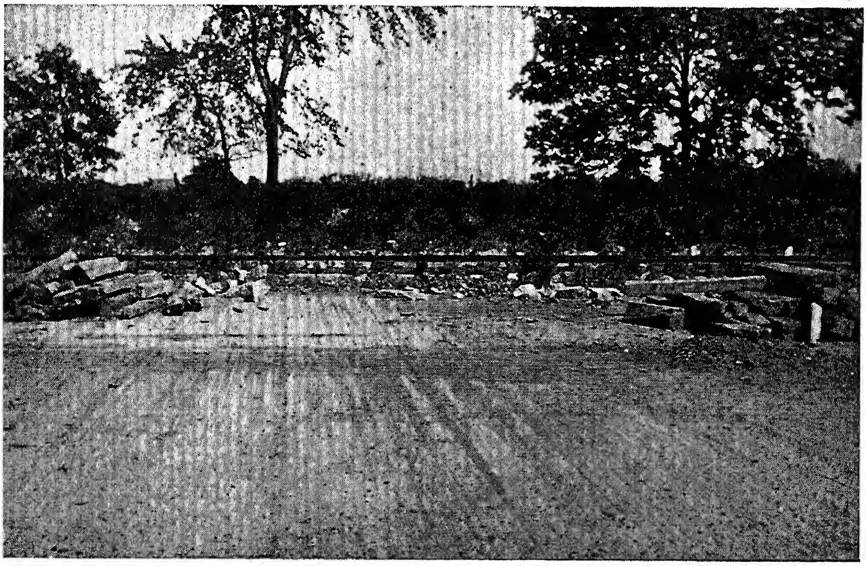

FIG. 197.—Glacial striae on bedrock, Kelleys Island, Ohio.

The effect of a continental ice sheet on the regional topography and relief is less obvious than that of a mountain or valley glacier. An ice sheet erodes the region between the valleys nearly as intensely as in the stream channels. Where the direction of movement of the glacier paralleled that of the major streams of the region, however, some of the valleys were gouged and deepened, and the regional relief was markedly increased.

Fig. 198.—Glacial grooves in bedrock, Kelleys Island, Ohio.

Fig. 199.—Unsorted glacial till.

The relief was decreased where the ice moved across the valleys and partially or completely filled them with glacial drift.

Depositional Features. *General Characteristics of Drift.*—Because of its origin, the most characteristic feature of the drift deposited by a

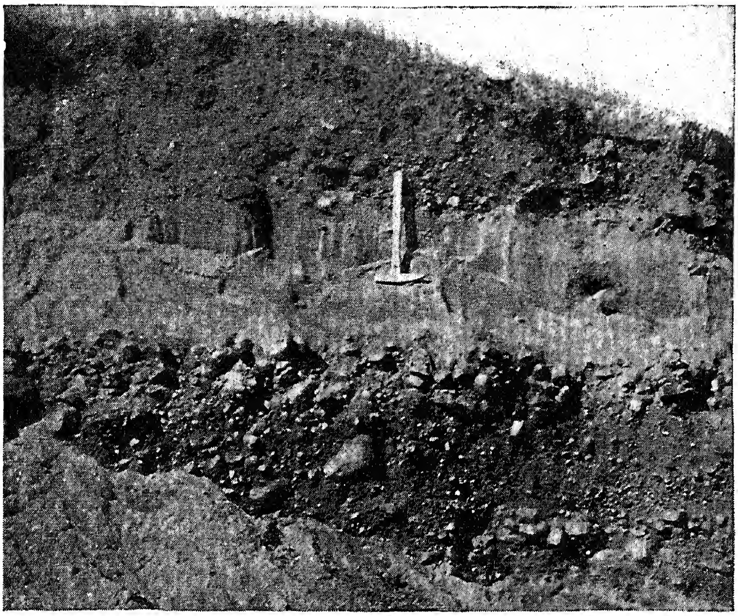

Fig. 200.—Interbedded sorted and unsorted drift.

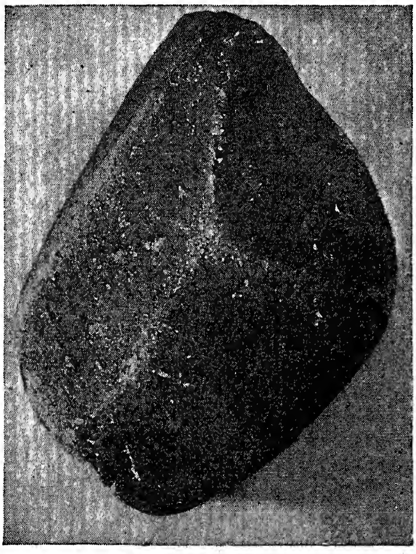

Fig. 201.

Fig. 202.

Fig. 201.—A faceted glacial pebble, showing the flattened surfaces produced by glacial brasion. (About one-half natural size.) These striae distinguish glacial pebbles from imilarly shaped "dreikanters" formed by wind abrasion.

Fig. 202.—Striated glacial cobble-stone from the drift at Norfolk, Nebraska. (*Photograph by Alden, U. S. Geol. Survey.*)

continental glacier is its heterogeneity (Fig. 199). In a small area may be found both stratified drift and unstratified till, with variations in size from fine clay and sand through coarse gravel (Fig. 200) to huge boulders. The size and shape of the particles are dependent to a considerable degree upon the resistance of the rock to abrasion (Figs. 201, 202). Rocks such as schists, clayey limestones, and shales are readily broken by the ice and are converted into glacial clay, silt, or small pebbles, whereas massive igneous rock, hard limestones, and quartzites are more resistant, and these constitute most of the larger boulders. Glacial drift generally contains rocks of many different kinds, depending upon those that are present in the areas over which the ice moved.

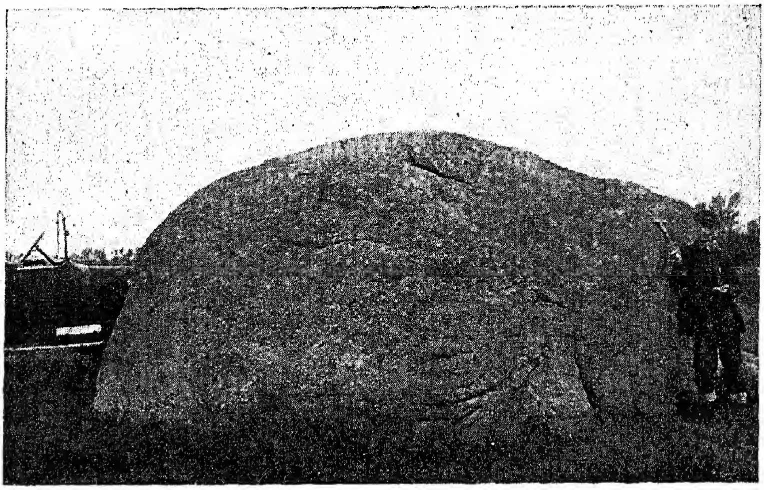

FIG. 203.—A glacial erratic. A huge granite boulder that is unlike underlying rock. Such boulders are found in positions where no agent of transportation other than ice could have placed them.

The large glacial boulders that are foreign to the underlying rock are *erratics* (Figs. 203, 204). Some immense masses of rock have traveled hundreds of miles from their sources. One of the largest erratics known in the United States is a mass of granite at Madison, New Hampshire, that measures 90 by 40 by 30 feet. Likewise on Mount Tom, near Northampton, Massachusetts, large, angular granite boulders are perched on the top of a high ridge of basic igneous rock. Such boulders are found in positions where no agent of transportation other than ice could have placed them.

The thickness of the drift varies greatly owing to (1) the irregularities of the surface upon which it was deposited, (2) the variations in the amount of débris carried by the ice, (3) the rate of advance or retreat of the glacier, and (4) the amount of erosion subsequent to its deposition. As a general rule, high, rugged regions have very thin coats of drift, and extensive low, smooth areas, like those of central Iowa and south central Minnesota, have from 1 to 600 feet of drift and very few rock outcrops.

Terminal Moraines.—The terminal moraines of a continental glacier are more conspicuous and more complex than those developed by a valley glacier. This is due in part to the absence of streams of high gradient, such as those which issue from mountain glaciers and transport much of the débris to the valley trains. Where the terminus of a continental glacier is in a partially peneplaned area, most of the morainic material remains where it was dropped at the margin of the melting ice. If the ice front is nearly stationary or fluctuates for an extended period in a zone not more than a few miles wide, series of morainal ridges are developed coinciding with the outline of the marginal lobes of the glacier, and these constitute the terminal moraines. Such moraines are characterized by many small, rounded hills of drift with adjacent depressions distributed in a disorderly fashion in a relatively long, narrow zone that paralleled the ice front at the time of deposition (Figs. 205, 206). In many regions one side of the system of hills has a steeper slope than the opposite side, and the steep slopes lie in the direction from which the ice came. Many of the larger ridges have superimposed upon them smaller hills and conical mounds, which make the depressions seem still deeper. Because of its irregularity such topography is commonly referred to as "knob-and-kettle," or "hummocky," topography (Fig. 207).

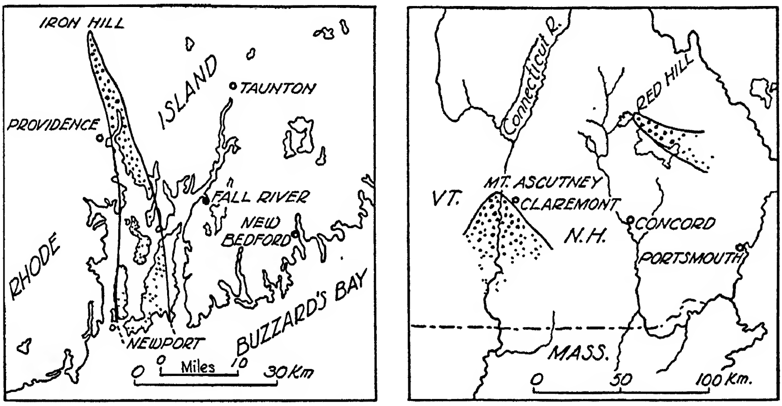

FIG. 204.—Maps showing glacial boulder trains in New England. The black iron-bearing rock of Iron Hill was distributed over a progressively wider area southward toward Newport. The rocks of Red Hill and Mt. Ascutney were spread in a similar manner.

Where a retreating glacier halts in its retreat and its edge remains in nearly a constant position for a considerable period of time, a moraine may develop. Such a deposit is a *recessional moraine* (Figs. 192, 206).

Ground Moraines.—The ground moraine is the most important and the most widespread depositional feature of a continental ice sheet.

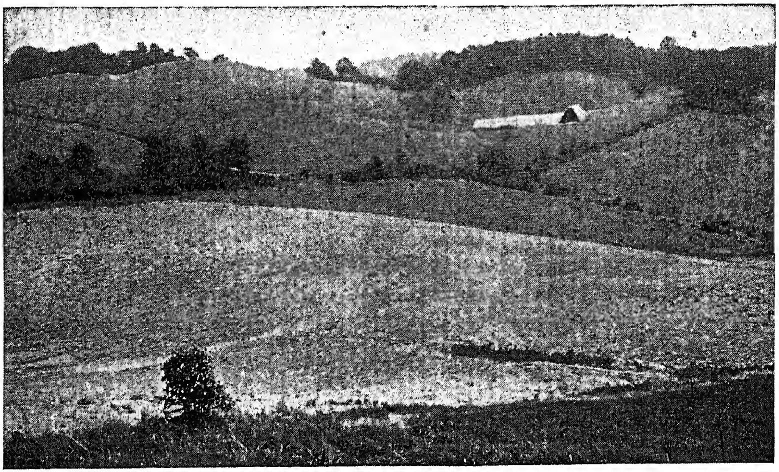

Fig. 205.—Glacial moraine topography near Saint Paul, Minnesota.

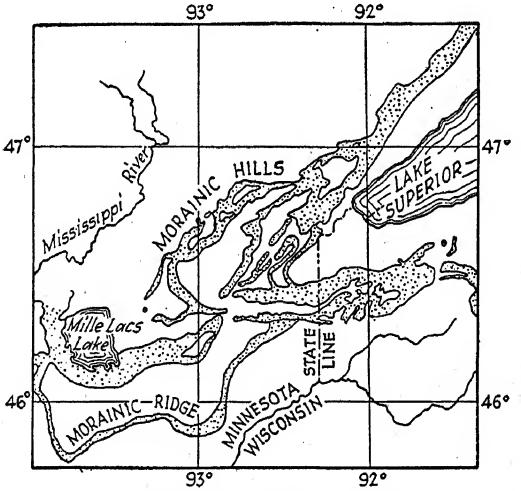

Fig. 206.—A group of morainic hills and ridges southwest of Lake Superior. (*Leverett, U. S. Geol. Survey.*)

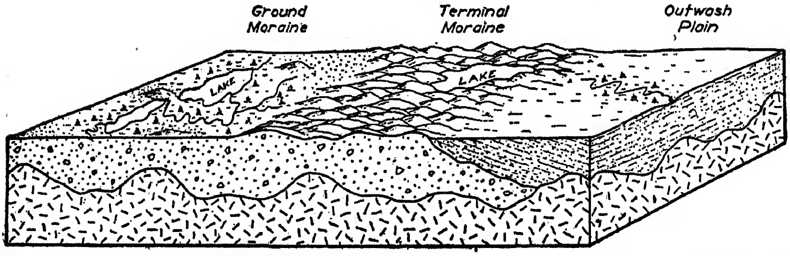

Fig. 207.—Diagram showing the field relations of an outwash plain, terminal moraine, nd ground moraine of a continental ice sheet. Note the rugged topography of the erminal moraine, the swamps and lakes of the poorly drained ground moraine, and the ow relief of the outwash plain.

It is the drift that was deposited below the ice during the period of recession. Instead of being concentrated in high knobs and ridges, as in the terminal moraine, it is scattered over the surface where the ice melted. Youthful ground-moraine topography commonly is characterized by numerous large depressions occupied by bogs, swamps, and lakes with no apparent relation between major streams and higher land.

FIG. 208.—A drumlin south of Newark, New York. The view is southwest. (*Photograph by Gilbert, U. S. Geol. Survey.*)

Such areas are usually poorly drained because of the damming of the preglacial streams and the unequal distribution of glacial till.

Certain areas of the ground moraine have small, smooth, oval hills of till that are lenticular in horizontal section and have their longer axes parallel to the movement of the ice which formerly covered them. Such hills are called *drumlins* (Fig. 208). They are commonly from 25 to 150 feet high. Many of them are 1,000 to 3,000 feet long and about 500 feet wide. In some regions they show many variations in size and shape from mammillary or dome-shaped hills to slender or linear ridges (Figs. 209, 210). They are conspicuously developed in the region around

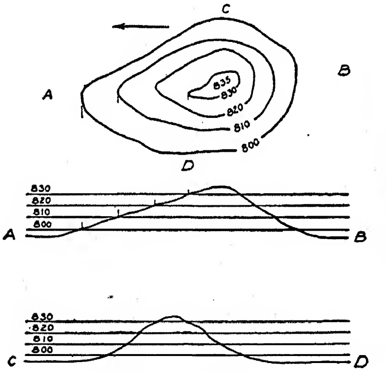

FIG. 209.

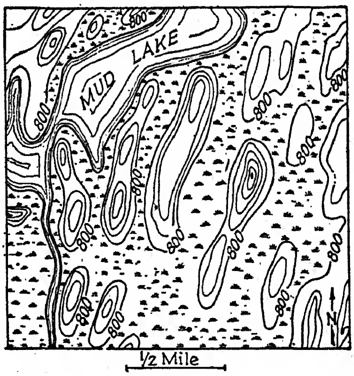

FIG. 210.

FIG. 209.—Plan and profile views of a drumlin. The arrow indicates direction of movement of the ice. The gentle slope commonly lies in the direction toward which the ice moved.

FIG. 210.—Drumlins near Waterloo, Wisconsin. (*U. S. Geol. Survey.*)

Boston, western New York, eastern Wisconsin, and north central Minnesota. Most drumlins are composed of unsorted till that exhibits very

FIG. 211.—An esker, Westford Township, Dodge County, Wisconsin. (*Photograph by Alden, U. S. Geol. Survey.*)

FIG. 212.—An esker or "serpent ridge" near Fort Ripley, Minnesotá, as seen from the air. (*Courtesy Dr. W. S. Cooper.*)

little lamination due to water action. The gentle slope of the drumlin lies in the direction toward which the ice moved. This is the opposite of the relation of the gentle slope to direction of movement where the ice erodes.

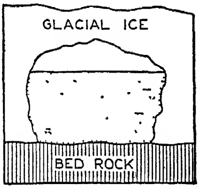

Fig. 213.—Diagram illustrating the origin of an esker. The ice tunnel containing a subglacial stream has built up its bed with sand and gravel.

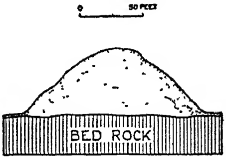

Fig. 214.—Cross section of the same stream deposit shown in Fig. 213, after the recession of the glacier and the melting of the supporting ice walls.

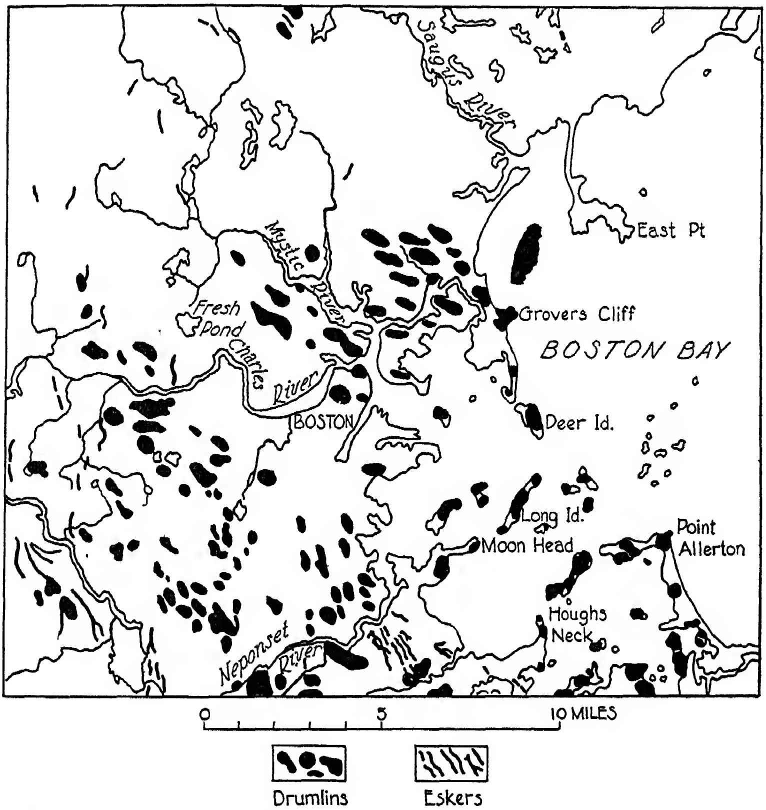

Fig. 215.—Map showing distribution and orientation of eskers and drumlins near Boston, Massachusetts. (*U. S. Geol. Survey.*)

Glaciofluviatile Deposits. *Eskers.*—Eskers are winding ridges (serpent ridges) of irregularly stratified sand and gravel that are found within the area of the ground moraine. Many of them are several miles long, and they are rarely more than a few rods wide, and their courses are roughly parallel to the direction of the movement of the glacier. Some are so nearly symmetrical in outline that they resemble railroad grades (Fig. 211). The ridges are evidently the beds of streams which flowed in tunnels or ice-walled gorges in or beneath the ice and aggraded their beds before the stream issued from the ice front (Figs. 212 to 215). Some undoubtedly are crevasse fillings. They are seldom continuous for long distances, and some of them serve as ridges or divides between shallow lakes and bogs in the ground moraine. Some of the hills of gravel deposited by heavily laden glacial streams flowing in and on the ice near the terminus are conical in outline and are referred to as *kames* (Figs. 216, 217).

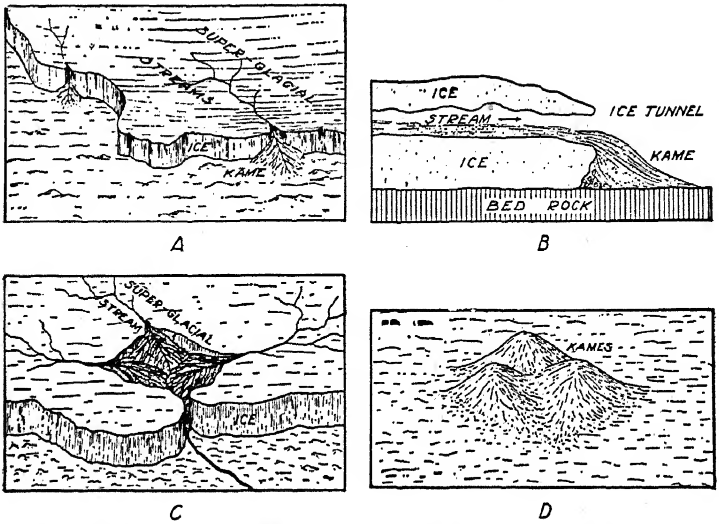

Fig. 216.—Diagrams illustrating the origin of kames. *A*, by deposition from superglacial streams at the margin of the ice; *B*, by englacial streams discharging at the margin of the glacier; *C*, by deposition from superglacial streams in deep reentrants between lobes; *D*, a group of kames after the ice has melted.

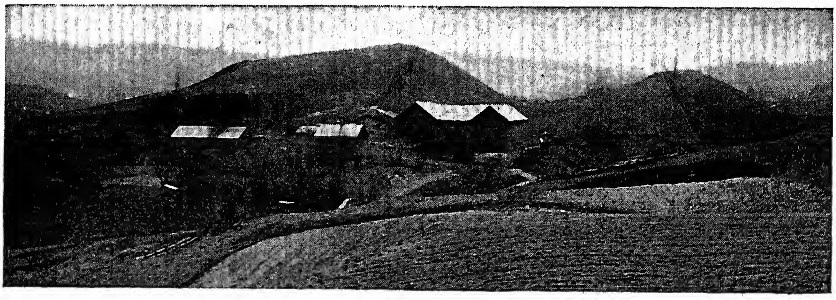

Fig. 217.—Kames in Ninemile Valley, New York. *(Photograph by Gilbert, U. S. Geol. Survey.)*

Outwash Plains.—The water from the melting ice at the edge of an ice sheet flows through the terminal moraine débris as a great number of streams rather than as a continuous sheet of water. Each of these streams builds a low alluvial fan, and the fans coalesce into a plain that slopes gently away from the terminal moraine area. Such a plain is composed of material washed out beyond the terminal moraine, and it is called an *outwash plain.* The heaviest load is deposited near the terminal moraine, and there the deposits often are composed of gravel and coarse sand, whereas farther away the slopes are more gentle and the deposits are fine sands and clay (Fig. 218). Many of the outwash plains of the northeastern part of the United States are so flat that often they are referred to as prairies.

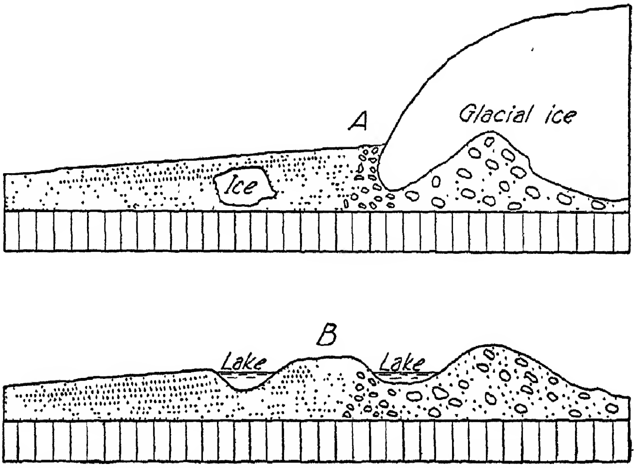

FIG. 218.—Diagrams showing the mode of origin of the outwash plain and its relation to the terminal moraine. At the margin of the ice, glacial waters build up a gently sloping plain composed of stratified sands. The terminal morainic debris is partly submarginal. When the terminus of the glacier recedes, a depression frequently is left between the terminal moraine and the outwash plain. This later becomes the site of a glacial lake or swamp. *A*, before melting of ice; *B*, after melting of ice. (*Modified after Woodworth.*)

Small kettle-like depressions with no outlets are formed in outwash plains by the melting of great masses of ice left during the recession of the ice front. Where such depressions are numerous, the outwash areas are *pitted plains.*

Temporary Glacial Lakes.—Since a continental glacier covers all of the land, it follows that divides between drainage basins also are buried under the ice. If a land surface in front of the glacier slopes toward the ice edge, water derived from the melting of the glacier will accumulate against the high land or divide, and the margin of the ice which serves as a dam will prevent the water from following its former course (Fig. 219). In this way large areas of land may be flooded forming marginal glacial lakes. Such lakes will exist until an outlet is cut across the lowest point in the divide. Numerous lakes were so formed along the margins of the continental ice sheets that covered parts of Europe and North America during the recent ice age. The outlets of many of these lakes cut broad and deep channels through the divides of drainage systems and at a number of places excavated valleys across continental divides where no streams exist at the present time. Most of these lakes were small and were soon filled with sediments derived from the glacier. Some, however, covered hundreds of square miles and existed for sufficiently long periods to form important depositional features.

The former presence of temporary glacial lakes is shown by shore-line features such as beaches, beach ridges, bars, and also by deltas and other lake sediments. At many places a series of shore lines are found at different elevations, due to fluctuating levels of the water.

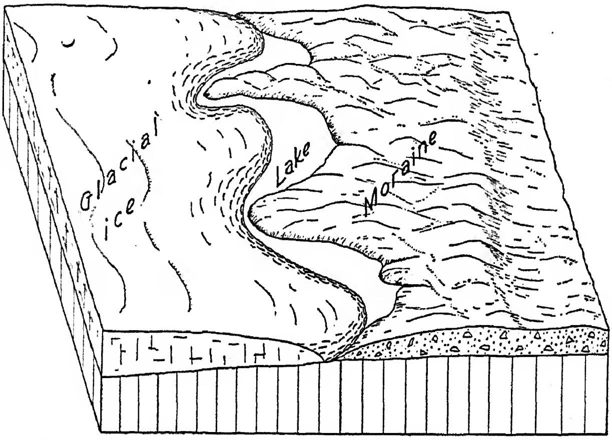

FIG. 219.—Diagrammatic cross section showing the origin of a marginal glacial lake. The normal preglacial drainage was to the left. The ice sheet served as a dam, blocking the flow of water in that direction.

Glacial Lake Agassiz.—The receding ice sheet southwest of Hudson Bay made the largest temporary glacial lake that was formed on the North American continent. Its development was due to the northward slope of the valley of the Red River of the North. After the glacier receded over the divide between the Minnesota River and the Red River, the water accumulated along the south and west margins of the ice, and, with a continued retreat of the ice, the lake increased in size and depth until eventually the water stood at a sufficiently high level to flow over the crest of the continental divide and into the Minnesota River at Browns Valley, Minnesota. When at its maximum extent, Lake Agassiz was about 700 miles long and 250 miles wide. It covered over 100,000 square miles, more than the combined area of all the present Great Lakes. Lake Winnipeg in Canada and Lake of the Woods on the international boundary occupy depressions in the bed of this ancient lake (Fig. 220).

Glacial Lake Sediments.—The cold waters of marginal glacial lakes have a higher density than water at moderate temperatures. For this reason the settling of silts and clays is retarded, and thus the fine sediments in suspension become diffused throughout the lake waters and are deposited over the floor of the entire basin. The fine-grained sediments are laid down in plainly separated annual layers called *varves.* Varve is a Swedish word that means the deposit of a season, whether of winter and summer or wet and dry seasons. Glacial lake varves commonly consist of two laminae, one of which was laid down during the summer and the other during the winter. The summer band consists of a light-colored, coarse silt, whereas the winter band is darker in color, finer grained, and thinner. The winter band is sharply separated from the summer band above but grades into the summer band below. Thus each varve represents a year. The varves vary in thickness and in other characteristics, and accordingly it is possible to match or correlate the top sets in one lake with the bottom sets in the next lake in the direction of recession of the ice and thus count the years consecutively. In this way the rate of recession of the last ice sheet has been calculated. It has been shown, for example, that 4,300 years elapsed while the ice was retreating 185 miles up the Connecticut River Valley from Hartford,

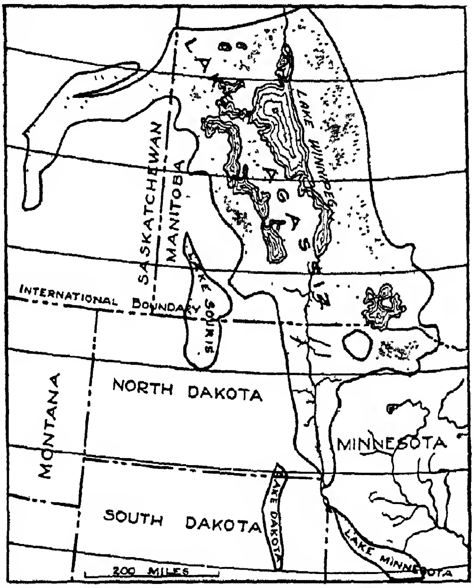

FIG. 220.—Map showing the boundaries of glacial Lake Agassiz during its highest water stage. (*After Upham, U. S. Geol. Survey.*)

Connecticut, to St. Johnsbury, Vermont. Varved clays are well exposed along Mink Brook at Hanover, New Hampshire.

Deposition from Stagnant Ice.—During the period of retreat of the last continental ice sheet, large areas of ice lost their forward motion and became great masses of stagnant ice. Their loss of motion may have been due to climatic changes or to the nature of the topography over which they stood. As the stagnant ice melted, no recessional or marginal moraines were formed, since such moraines can form only where there is forward movement of the ice accompanied by melting at the ice front. Where wastage takes place without movement, stratified drift is more abundant than glacial till. The preponderance of stratified drift is due to the fact that melting took place over a large area instead of being confined for the most part to the marginal zone of the glacier. Under such conditions, melt-water streams on and under the ice do not have their channels disturbed by the movement of the ice. Consequently the water flowing through crevasses and tunnels in the stagnant ice builds up long, narrow strings or trains of sand and gravel which stand as ridges of stratified drift after the ice has wasted away. Many eskers were built in this way. Where the ice had an intersecting network of cracks and crevasses that became partially filled with sediments, the *crevasse fillings* formed ridges or hummocks that enclosed depressions, or *kettles*, when the blocks of ice between the crevasses melted away. Many of the kettles now contain ponds, lakes, or swamps. Where lakes fed by melt-water existed at the margin or on the surface of the stagnant ice, streams deposited deltas and lake sediments on the floors of the lakes. When the ice was melted and the water level lowered, such lake sediments remained as flat-topped terraces along the channels through which the waters drained.

Stream Diversion through Glacial Deposition.—The unequal deposition of glacial drift had a profound effect upon the preglacial topography of the areas that were glaciated (Fig. 221). Numerous valleys were completely filled with glacial till, and as the ice receded many old channels were so obstructed that new surface-drainage courses were established (Fig. 222). Even such major streams as the St. Lawrence, the Mississippi, the Ohio, and the Missouri rivers locally were turned from their preexisting channels by the work of the glaciers. In regions of rugged and mountainous topography the channels have not been so greatly diverted, but in a number of regions drift accumulations in a valley buried projecting spurs of bedrock, and postglacial streams eroding new channels through such deposits have become superimposed on the old buried ledges that formerly were parts of old valley walls.

Ancient Glacial Deposits.—A geologically ancient glacial till which has been cemented into a firm rock is called *tillite.* Such ancient tillites

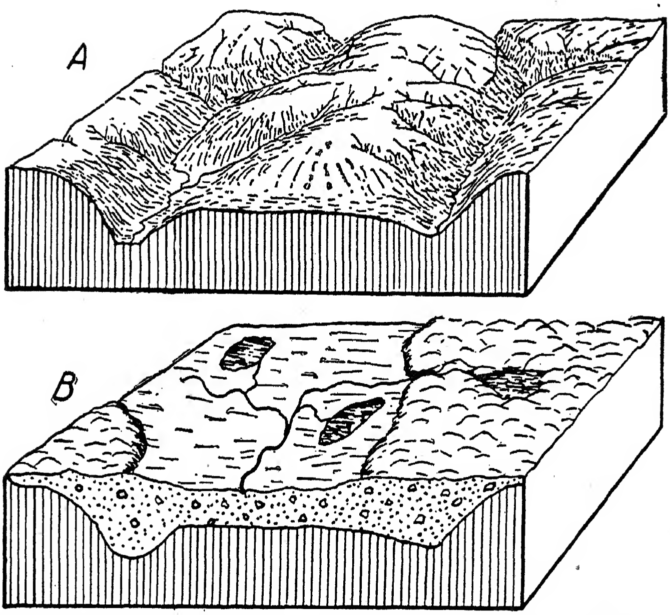

FIG. 221.—Diagrams illustrating topographic changes produced by glaciation. *A*, region maturely dissected by streams; *B*, same region after glaciation.

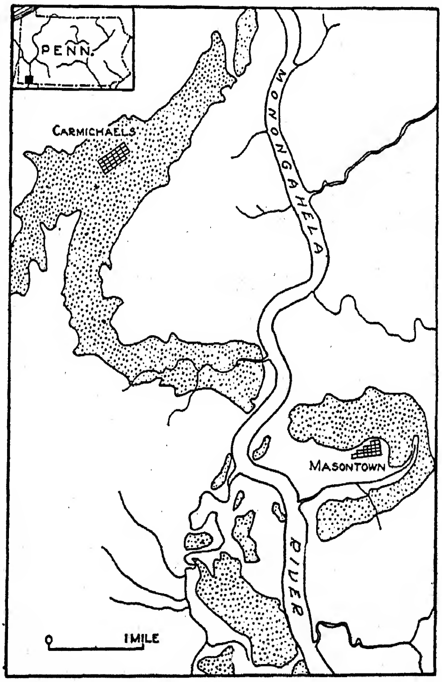

FIG. 222.—Drainage changes due to glaciation. The stippled zone shows the location of the preglacial valley. (*From map by Campbell, U. S. Geol. Survey.*)

have been recognized at many places, and they are supposed to be the deposits of ice sheets which covered the regions in former geological periods. Near Kimberly, South Africa, a bed of hard tillite occurs over glaciated bedrock surfaces that are striated and polished. In Australia thick tillite beds occur at several geological horizons. Similar, even more ancient deposits have been reported from central Canada and from northern Norway. In eastern Massachusetts a Permian conglomerate is regarded as tillite or glacial conglomerate. The formation is exposed at many places in the Boston Basin where it contains striated and faceted pebbles and other indications of its glacial origin.

Causes of Glaciation.—Many hypotheses have been offered to account for the climate which resulted in continental glaciation, but none is generally accepted. One possible cause is a variation in solar energy due to the variation in the eccentricity of the earth's orbit. Another theory postulates great elevation of the land in the areas that were glaciated. A more recent atmospheric hypothesis, developed by T. C. Chamberlin, is based upon possible variations in the amount of carbon dioxide and water vapor in the atmosphere. Experimental evidence indicates that both carbon dioxide and water prevent the radiation of much of the heat derived from the sun. When these substances are less abundant, the amount of radiation increases, and a colder climate ensues. The enlarging of the land areas may have decreased the amount of water vapor in the atmosphere and thereby decreased its ability to retain the heat derived from the sun.

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CHAPTER VII

LAKES AND MARSHES

A lake is a natural inland depression or reservoir containing an appreciable amount of water. Such bodies of water vary in size from small ponds to great inland seas covering many thousands of square miles, such as the Caspian Sea or the Great Lakes. They are widely distributed over the earth's surface, occurring in mountain areas, on plateaus and plains, along valleys and sea coasts. They vary in depth from shallow water covering low, boggy depressions to rocky basins over 5,000 feet deep. Soundings in Lake Baikal in Siberia have recorded a depth of 5,618 feet. Lake waters vary in chemical composition from the soft, fresh water of a mountain lake fed by glaciers to the bitter, concentrated waters of the Great Salt Lake or the Dead Sea. Their composition depends largely upon the composition of the rocks over which and through which the waters that feed them have passed, and the extent to which they have been concentrated.

Most lakes have outlet streams that feed rivers. Such lake basins act as safety valves for the rivers they feed, for they tend to regulate the volume of discharge and thereby to prevent floods. The flood waters of inflowing streams spread out over the wide basin of a lake and raise the water level so slowly as to cause but slight damage along the valley of the main outlet. In like manner, during seasons of drought, the water from lakes flows out more slowly, and thus the lakes aid in preserving the permanent streams. Lakes also exert an influence upon local climatic conditions by increasing the rainfall. They tend to temper both the cold of winter and the heat of summer. They serve likewise as settling basins in the drainage systems of the land. Since they receive the sediments transported by the inflowing streams and the débris resulting from the action of waves and organic life along their shores, they become filled so that their duration in terms of geologic time is short.

Origin of Lake Basins.—Any geologic process which may produce a depression upon the surface of the earth or obstruct drainage channels may produce a lake. Most lake basins are the results of gradational processes, but some are due to diastrophic earth movements, and others are due to volcanic activity.

Basins Due to Glaciation.—Lakes generally are numerous in areas recently covered with glacial ice. Some glacial lakes are due to erosion

(Figs. 223, 224), others to deposition, and still others to a combination of these two processes. Numerous lake basins in mountain valleys were dug out of the mountain slopes by glacial abrasion. Such lakes occupy rock basins which are either smoothed and striated or are walled by jagged cliffs formed by glacial plucking. Cirque lakes are perhaps the most common type in mountainous regions. Valley glaciers in their retreat frequently form crescent-shaped ridges of débris, convex down-stream, which serve as dams and retain lakes. Many valleys of the Cordilleras contain lakes that were so formed. Numerous examples may be seen in the Wasatch mountains of Utah. Glaciers also may obstruct the drainage of lateral valleys so as to cause lakes to form.

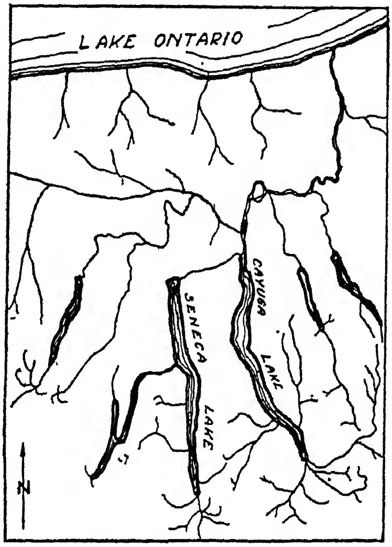

Fig. 223.—Map showing the shape and orientation of the Finger Lakes of New York. The lake basins represent preglacial valleys that have been deepened by glacial erosion.

Depressions on Drift Sheets.—Lake basins are abundant on the surface of the drift sheets that cover northeastern North America and northwestern Europe. In these regions hundreds of thousands of lake basins were formed. These vary in size from small ponds to basins that cover over 20,000 square miles. In Michigan, Wisconsin, and Minnesota, where the drift surface is unusually rugged at places, the lakes number more than a score to the square mile. Some occur in shallow basins on the outwash plains; others are surrounded by extensive marshes on the ground moraine; and the deepest and most irregular basins are found in the terminal moraine zones (Fig. 225).

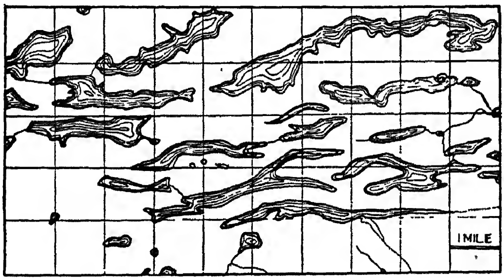

Fig. 224.—Parallel rock-basin lakes due to differential glacial erosion. The long axes of the lakes parallel the strikes of steeply dipping slaty rocks. Lake County, northeastern Minnesota.

Basins Formed by Streams.—Lake basins are formed by both the degradational and the aggradational work of streams. At certain places the two processes have been active in the formation of a single depression. Many shallow lakes are formed upon flood plains by streams that wander in meandering loops over the valley flats. Where the stream cuts off a meander, the abandoned part of the channel remains as a crescent-shaped *oxbow lake,* or *bayou* (Fig. 226). Numerous irregular lake basins are formed also in the depressions between the natural levees and the outer edges of the flood plains. These commonly are referred to as lateral lakes. Still farther down the valley, on the delta surface, the waters in the main channels break through the levees and form branching channels, which in turn form still other distributaries. In this way shallow depressions are completely surrounded by delta sediments, and the basins are converted into *delta lakes* (Fig. 227). The deltas of the Nile, Danube, Ganges, and Mississippi rivers show such basins now in the process of construction. Lake Borgne in Louisiana is not yet completed. The low delta lands of Holland also contain many lakes of this type.

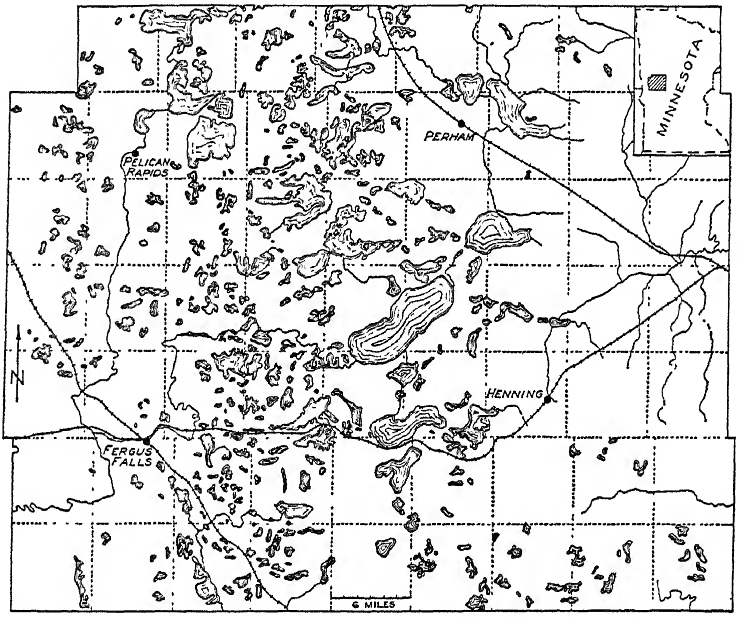

FIG. 225.—Irregular lakes in glacial-drift basins. West central Minnesota.

Where a tributary stream brings to the main stream an excess of sediment, it is deposited as an obstruction or dam in the channel of the main stream, and the water is ponded, forming a river lake. This is illustrated strikingly in the Mississippi River above the mouth of the Chippewa River of Wisconsin, where the Mississippi expands over its flood plain and forms Lake Pepin (Fig. 228). In the great valley of California the high gradient Kings River that flows from the Sierra Nevada Range has built an alluvial fan that obstructs the drainage to such an extent that Tulare Lake is formed above the fan. Lake San Cristobal in Colorado owes its origin to the Slumgullion mud flow, the lower part of which has built a dam across a valley (Fig. 229).

The lodgment of driftwood in the bed of a low-gradient stream may pond the water and form a lake in the stream valley. These rafts or jams of trees and logs are formed of timber that caves into the stream from forested banks that are undercut. A remarkable series of irregular lakes has been formed by this process along the Red River in Louisiana (Fig. 230), where timber rafts several square miles in area, covered with living vegetation, formed floating islands and jams that dammed the streams so as to cause the development of shallow lakes 20 to 30 miles long. A number of the jams have been removed, and the main stream has deepened its channel as much as 15 feet. As a result

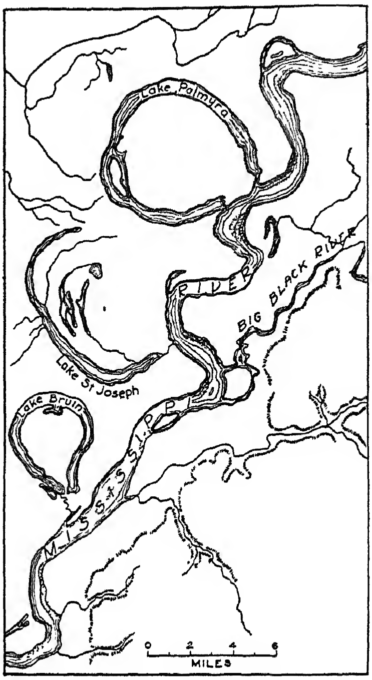

Fig. 226.—Oxbow lakes along the valley of the Mississippi River in Louisiana. (*After Russell.*)

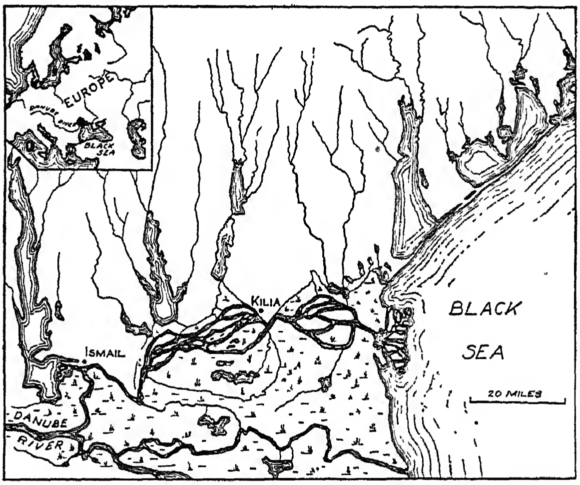

Fig. 227.—Lakes formed by delta sediments damming the outlets of tributary streams. (*After Reclus.*)

of this deepening, the tributaries have lowered their valleys, and the lakes are being drained.

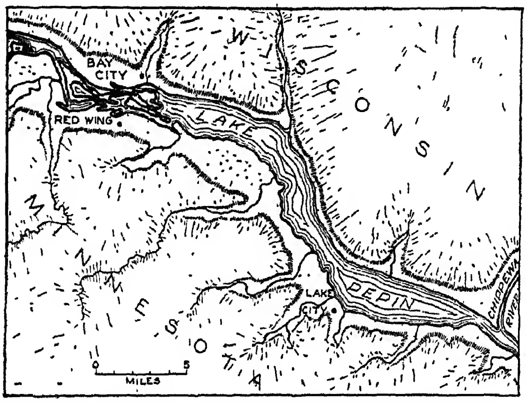

FIG. 228.—Lake Pepin, part of the Mississippi River where it has spread out over its former flood plain. The widening of the river is due to the deposition of an excess of sediment brought to the valley by the Chippewa River.

Streams may excavate depressions in their beds that later form lake basins, particularly at the bases of rapids and waterfalls, where corrasion is accelerated. Many examples may be seen in the channels of mountain streams and at the bases of waterfalls such as those of Niagara. Where stream courses with such "plunge basins" are abandoned, the depressions

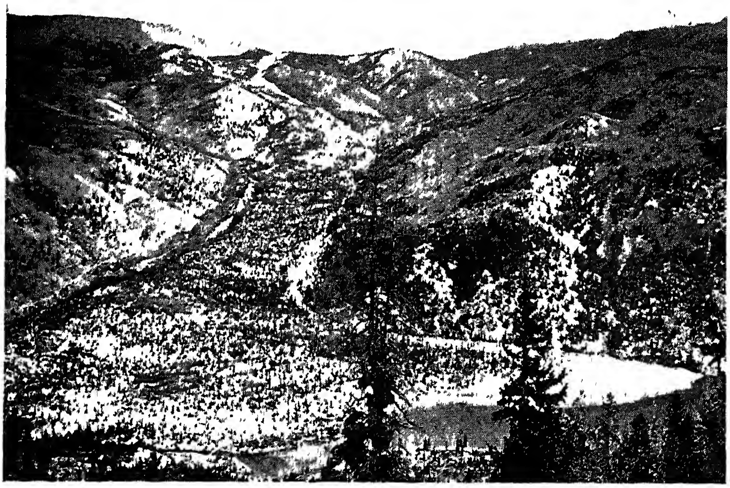

FIG. 229.—Slumgullion mud flow at Lake San Cristobal, Colorado. The lower part of the mud flow acts as a dam, forming a lake.

become lakes. Grand Coulee Lake in the state of Washington and Jamesville Lake near Syracuse, New York, are examples.

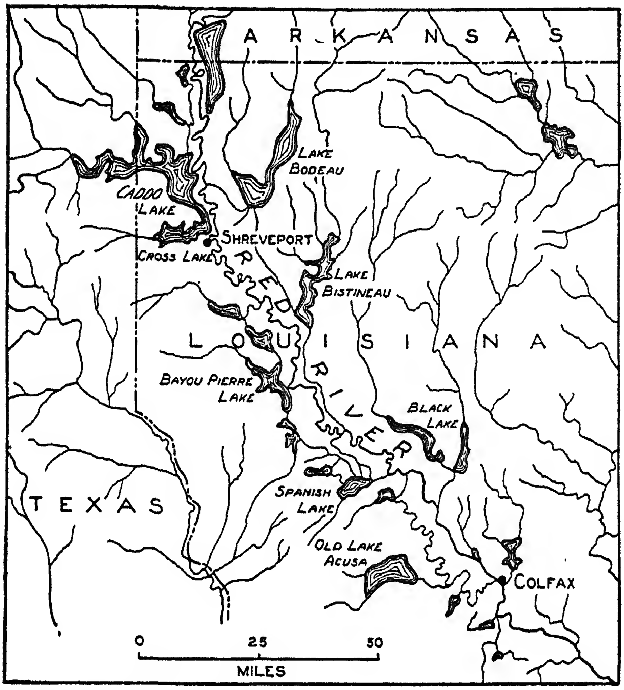

FIG. 230.—Valley lakes along the Red River in Louisiana. Some of the lakes were formed by log rafts acting as dams. (*After Veatch, U. S. Geol. Survey.*)

Subsurface streams also may lead to the formation of lake basins at the surface. It has been pointed out that the solution and removal by underground water of large quantities of carbonates of lime and magnesia from regions underlain by limestone lead to the formation of underground cavities and channels. As such cavities increase in size, their roofs may collapse, and depressions are formed at the surface. If the

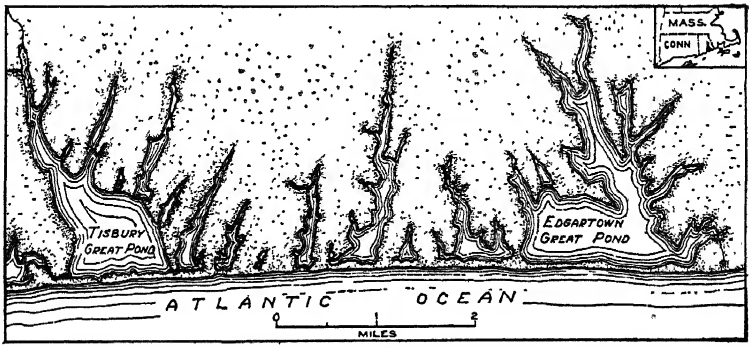

FIG. 231.—Coastal-plain lakes due to wave action. Marthas Vineyard islands, Massachusetts. (*U. S. Geol. Survey.*)

regional ground-water level later is brought near to the surface, the basins are occupied by lakes. A number of lake basins in Florida were formed in this way, and other examples are found in the limestone sink, or "karst," regions of Tennessee and Kentucky.

Basins Formed by Waves and Shore Currents.—The combined action of wind and waves along the shores of the oceans, seas, and large lakes leads to the formation of sand and gravel bars, forming lakes and lagoons that are landlocked. They are most numerous where bars have been built across the entrances of bays or where they extend from a headland and cut off a curve of the original shore line. Many such lakes occur along the shore of Lake Ontario and along the Atlantic Coast (Fig. 231) from New York southwest to Panama.

Basins Due to Diastrophism.—The crust of the earth is unstable, and movements of land areas have resulted in the formation of depressions that retain water. Folding produced an enormous trough-like basin that contributed to the formation of Lake Superior. The warping of a river valley may likewise produce a lake basin. This is true especially if the warping elevates one part of the valley so that it is higher than the part of the valley toward the source of the stream. In such a basin the water is ponded and forms a lake. Some large lake basins have been formed by the uplift of mountains around them. This is shown on a large scale in the Great Basin area of Utah and adjoining states.

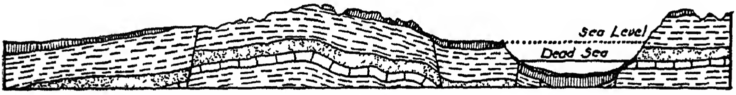

FIG. 232.—Section across the Dead Sea showing the outline of its basin, which was produced by the depression of a block of the earth's crust along fractures or faults. (*After Blanckenhorn.*)

Warping and folding often are accompanied by the breaking and displacement of rock strata along faults. Differential movement along fault blocks may cause part of a valley to sink, thus giving rise to a lake basin. A chain of valleys due to downfaulting of narrow blocks of the earth's crust have formed the Great Rift Valley in Asia and Africa. This rift includes the Jordan River, the Dead Sea (Fig. 232), part of the Red Sea, the upper Nile, and the chain of African lakes including Tanganyika, Leopold, and Nyassa. The basin of Lake Tanganyika extends 1,700 feet below sea level. In the western part of the United States a number of basins have been formed by the same process. Lake Tahoe on the California-Nevada state boundary is one of the deepest lakes in North America. There an earth block settled several thousand feet to form a lake basin 22 miles long and 12 miles wide. The basins of Warren Lake and of Abert Lake in Oregon also are due to faulting.

Undulations of the earth's surface during earthquakes may produce depressions that become filled with water and form lakes. Reelfoot Lake in western Tennessee was formed during the earthquake of 1811, where there was sinking of numerous areas in Arkansas, Tennessee, and Missouri.

Where diastrophism causes a recession of the sea, the newly emerged portions of the sea bottom may contain basin-like depressions that remain as lake basins. A number of lakes in Florida are examples. The lakes are surrounded by marine sediments of recent origin, all of which are but a few feet above sea level. Such lakes are in regions of extreme topographic youth, and when drainage channels are established the shallow basins soon are emptied.

FIG. 233.—Crater Lake, Oregon. Wizard Island near the opposite shore is a small volcanic cone within the crater. (*Courtesy Sawyer Scenic Photos, Inc.*)

Basins Produced by Volcanic Activity.—Lava, rising from the earth's interior through vents in the surface, may fill or so obstruct a river valley that a lake is formed. Such basins once existed on the west slope of the Sierra Nevada, but the lakes have been drained by the erosion of the lava dams.

Lakes also occupy the hollows of extinct volcanic craters, some of which are at considerable altitudes above the surrounding country. In the western part of the United States examples are found in Arizona, Nevada, California, and Oregon. Of this group Crater Lake in Oregon (Fig. 233) is the most renowned. This lake shows such extraordinary geological features that the region it occupies has been declared a national park. Crater Lake is situated in the Cascade Mountains in southwestern Oregon, at an elevation of 6,177 feet above the sea. It is nearly circular, without bays or promontories, and is about 5 miles in diameter (Fig. 234). The cliffs of dark andesite encircling it rise precipitously to heights varying from 900 to 2,200 feet and nowhere offer an easy means of access to the basin within. There are no streams tributary to the lake, and no visible outlet. The waters probably escape by seeping through the walls and floor, as precipitation in this region is in excess of evaporation.

FIG. 234.—Photograph of a model of the basin of Crater Lake. (*U. S. Geol. Survey.*)

The basin of Crater Lake presumably was formed by the collapse and subsidence of the core of the volcanic mountain, carrying the summit downward as it sank. This was perhaps due to the withdrawal of lava from the mass of the magma deep in the earth beneath the volcano. After the removal of the summit of the mountain, the volcano again became active and built a small volcanic cone within the ancient crater. This cone is Wizard Island in the present lake (Fig. 235).

Saline Lakes.—The waters of lakes vary much as to their mineral content. Those that have outlets generally are fresh. The waters that enter them generally have a low content of dissolved material. The salinity is increased somewhat by evaporation, but there is no great concentration of salts, because the water flowing out of the lake is renewed by fresh water. Lakes that are in undrained basins also are continually receiving fresh water of low mineral content. The waters are evaporated from such lakes; and although the water entering the lake may carry only small amounts of mineral matter, this is concentrated in the lakes. Since there is no outlet, the water steadily becomes more salty by evaporation, and it may reach the point where it becomes saturated and deposits salt. The kind of salt that predominates depends upon the chemical composition of the rocks in the region of the lake basin and subordinately upon the springs that rise in the basin. In some it is near the salinity

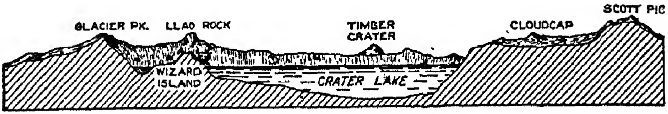

FIG. 235.—Cross section of the basin of Crater Lake, Oregon. Crater Lake is about 5 miles in diameter. (*After Diller, U. S. Geol. Survey.*)

of sea water; in others alkaline salts predominate; and in several localities borax salts are abundant.

Some salt lakes are of oceanic origin. The Caspian Sea, which is the largest body of inland salt water known, was cut off from the ocean by the elevation of the intervening land. The most favorable conditions for the formation of terrestrial saline lakes are in regions where mountain streams discharge into interior drainage basins where the climate is arid. Such conditions are found over a large area in the southwestern part of the United States (Fig. 236).

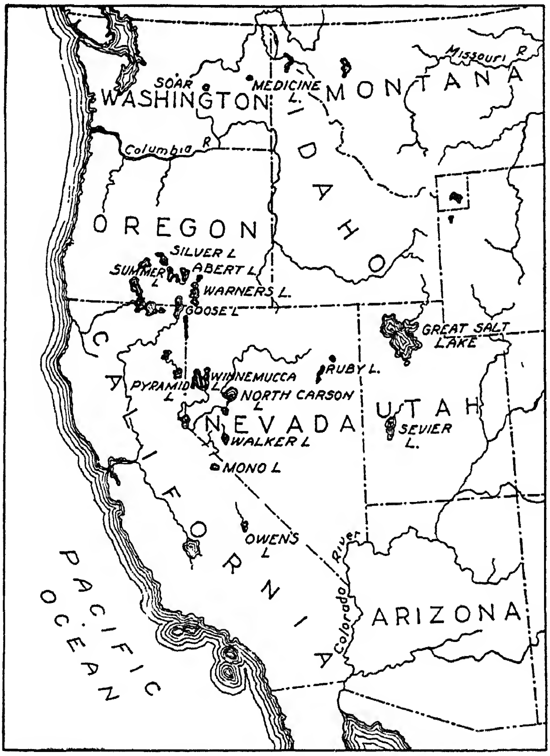

FIG. 236.—Map showing saline lakes in the western part of the United States. (*After Russell.*)

Great Salt Lake, Utah.—Great Salt Lake is but a shrunken remnant that contains the residue of a vast fresh-water lake that once covered an area of 20,000 square miles, mainly to the west and south of its present lake shores. The fresh-water predecessor, known as Lake Bonneville, overflowed to the north through the Snake and Columbia rivers into the Pacific Ocean (Fig. 237). It left a record of its boundaries in the form of hundreds of miles of shore beaches, terraces, and wave-cut cliffs which it formed along its mountainous shores at the different levels at which it stood (Fig. 238). Lake Bonneville once had a maximum depth

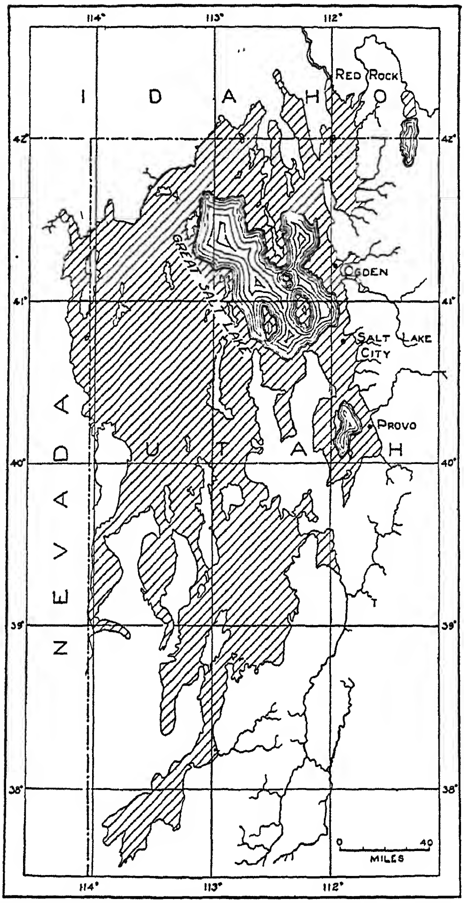

FIG. 237.—Map of Great Salt Lake showing its relation to the vast ancestral lake known as Lake Bonneville (shaded). (*After Gilbert, U. S. Geol. Survey.*)

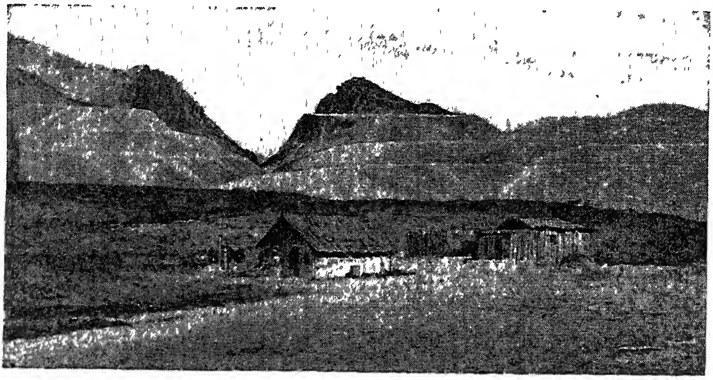

FIG. 238.—Ancient shore-line terraces of Lake Bonneville. (*After Gilbert, U. S. Geol. Survey.*)

of over 1,000 feet. Where the famous Mormon temple now stands in Salt Lake City the water of this lake was once so deep that the temple would be under 850 feet of water if the lake again attained its former level. As the level of Lake Bonneville sank below its outlet, because of diminished rainfall in that region, its waters became more and more salty, until the present Great Salt Lake is nearly five times as salty as the ocean. It has been estimated that the lake water contains 400,000,000 tons of common salt and also millions of tons of salts of calcium, magnesium, potash, and other constituents. In marginal pools, where the water is only a few inches deep, it becomes so concentrated by evaporation that common-salt crystals are formed and accumulate as a brilliant white sediment under the water.

Alkaline Lakes.—Lakes containing excessive amounts of alkaline carbonates are commonly called alkaline lakes. They occur at many places in Egypt, Hungary, Venezuela, etc. They are found along the western part of the Great Basin in the United States. Mono Lake in east central California is an example. Its basin lies at the eastern base of the Sierra Nevada, where it covers an area of about 90 square miles. Its size varies with the seasons and, to some extent, from year to year. This lake also, like many in the Great Basin country, had a much larger ancestor, as is shown by numerous shore features carved on the slopes of its basin. The highest beach line is nearly 700 feet above the present water level. When filled to that beach, it covered an area of about 320 square miles. The lakes in the region of Great Salt Lake are surrounded by marine sedimentary rocks that supply sodium chloride. Hot springs are numerous in the southwestern part of the United States and contribute large amounts of salts to certain lakes.

Playa Lakes.—Shallow, flat-bottomed depressions are formed by weathering and wind action in desert areas. During the wet seasons they are flooded by waters from intermittent streams; and when the water supply ceases, they shrink and finally disappear by evaporation, leaving in their beds deposits of alkali salts that often are as white as freshly fallen snow. Such ephemeral lakes are characteristic features in the deserts of the Great Basin. In the Black Rock desert in the northwestern part of Nevada during the winter months a lake forms that covers an area of 450 to 500 square miles and is seldom more than a few inches deep. Following heavy storms it appears as a vast sheet of liquid mud. In a few days all the water may evaporate, leaving a hard, dry, barren surface showing a reticulate pattern of mud cracks. Similar lakes exist in the deserts of Arizona, New Mexico, and Sonora. They are found also in the desert areas of other continents.

Processes Modifying Lake Basins.—The form of a lake basin depends upon the forces that operate at the time of its formation. Whatever the original outline may be, it does not long remain the same, for gradation is constantly at work modifying the lake shores, filling the basin with sediments, or draining it. Sooner or later most lakes are destroyed either by the lowering of their outlets or by the filling of their basins. At places their extinction is due to evaporation or to a diversion of inflowing waters.

The downward cutting of outlet streams lowers the water level of the lakes, and the depth of the basins is gradually diminished. This process has been an important factor in the destruction of many glacial lakes in eastern Canada and the northeastern part of the United States. Such lakes, with the rims of their basins composed of unconsolidated glacial débris, are drained more rapidly than basins in hard igneous or sedimentary rocks. The outlet stream of a lake is always a poor agent of erosion, for the quiet lake serves as a settling basin for the silts brought from the surrounding slopes. Therefore, the outlets carry such clear water that very little erosion is accomplished where they flow over massive rock formations.

The filling of lake basins takes place in various ways.

1. Waves are almost continually engaged in cutting into the lake shores, and the material worn away is deposited in their basins. On the west side of Lake Michigan, cliffs at places are receding at an average rate of 5 feet per year; and though such erosion enlarges the area of the lake basin, the water is made shallower by the materials eroded from above the water level.

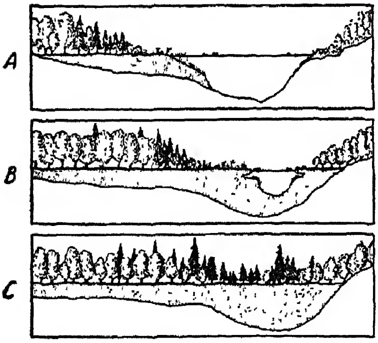

FIG. 239.—Diagrams showing the successive filling stages of lakes by peat. (*After Dachnowski, Ohio Geol. Survey.*)

2. Deltas are built where inflowing streams reach the quiet lake water. Many are built up above the water level, and the area of the lake is reduced. Several of the Finger Lakes of New York have been shortened as much as 3 miles by delta plains that have been built in their basins. A more striking example is Lake Geneva in Switzerland, where the Rhone River, fed by glacial streams charged with sediments, has built a delta 20 miles long into a lake basin that is nearly 1,000 feet deep. The bank-like outer margin of this delta is advancing further into the lake each year, and if the process continues long enough the basin will be completely filled with sediments.

3. Wind-blown sands may be carried out into lakes, especially in arid regions, and eolian sediments such as volcanic ash have accumulated to considerable thicknesses in certain lake basins.

4. Remains of animals and plants accumulate on the bottoms and along the margins of lakes. This is particularly well shown around small, shallow lakes where aquatic vegetation grows. The plants along the edge of the water catch and hold the sediments along the shore and thus build the shore farther and farther out into the lake (Fig. 239). A great number of animals and certain groups of plants secrete calcium carbonate or silica. Such shells and skeletal remains also contribute to the filling of lake basins.

Economic Aspects of Lakes and Swamps.—Natural products that are beneficial to mankind are constantly being taken from lake waters or from the sediments that accumulate under the water. Saline lakes furnish salts and many shallow lakes have extensive deposits of calcium carbonate in the form of a white, chalky marl. This calcium carbonate is used for agricultural purposes and also in the manufacture of cement. In some localities the remains of microscopic plants called diatoms, which secrete tiny shells of silica, accumulate to form beds of diatomacous earth in lake basins. This material is excavated and used in the manufacture of abrasives, refractories, and other products. Extensive deposits of peat have been built up along the shores of lakes and bogs. In many localities in Europe peat cut from the bogs is dried and used as a fuel. Where lake basins have been filled or drained, either by natural or by artificial means, the exposed sediments become fertile soils. Large areas of lake soils are being cultivated in the "wheat belt" of North Dakota and western Minnesota, where the plains represent the drained basin of the extinct Lake Agassiz. Numerous coastal-plain swamps also have been drained or filled with sediments, leaving fertile soils high in humus. This is especially true along the southeastern coast of the United States, in Holland, and in Flanders.

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CHAPTER VIII

THE OCEAN

Distribution.—The ocean is that great body of water which occupies the depressed basin-like portions of the earth's surface. It lies between and surrounds the elevated or continental portions of the earth (Figs. 240, 241). The partially separated water areas have different names. Thus there are the Atlantic Ocean, the Pacific Ocean, etc., somewhat independent but nevertheless connected at the surface and forming a common ocean.

The outlying divisions of the ocean, such as the Red, the Mediterranean, and the Caribbean, are commonly called seas, but a nearly similar body of salt water may be called a gulf, as the Gulf of Mexico, or a bay, as the Hudson Bay. The whole ocean, moreover, may be called the sea, and it is the usual practice to refer to its surface as sea level. Some of these divisions or seas are almost completely surrounded by land, whereas others are partly inclosed by island chains. They may be entirely within the shallow water or may lie in isolated portions of the great ocean basins themselves. The processes at work within the ocean are the same whether in one of these outlying divisions or in the open ocean.

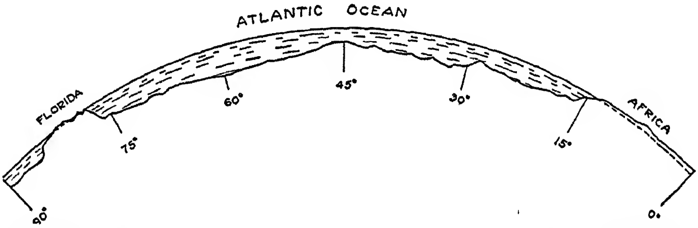

FIG. 240.—Profile of the Atlantic Ocean basin along the twenty-fifth parallel from Africa to Florida. (*Relief features after Murry and Hjort.*)

Functions of the Ocean.—The ocean supplies moisture to the atmosphere, from which it falls as rain on all parts of the earth's surface. It is the ultimate source of the land waters which are so effective in carving the continents and which give life to plants and animals. Its basin is the place toward which land detritus is being carried and is the final resting place of all sediments. The ocean absorbs heat slowly and gives it up slowly. It is therefore a great regulator of climate. Extending from continent to continent, it is the route over which marine life migrates, and it is the great highway of commerce, while its waters supply food for millions of people. From the geologic, the climatic,

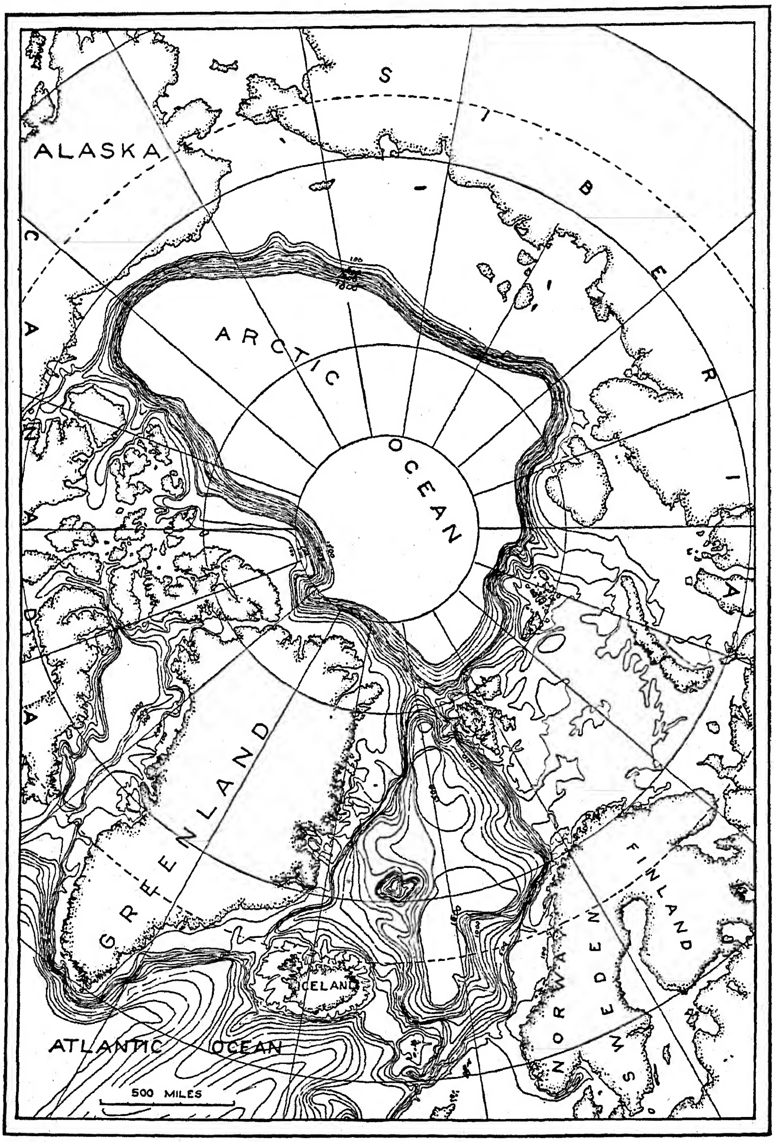

FIG. 241.—Map of the Arctic Ocean basin and submerged connections between North America and adjacent continents. Depths in fathoms. (*Modified from Nansen.*)

and the economic standpoint, therefore, the ocean is of vital, often dominating, importance, and its influence reaches every part of the earth's surface.

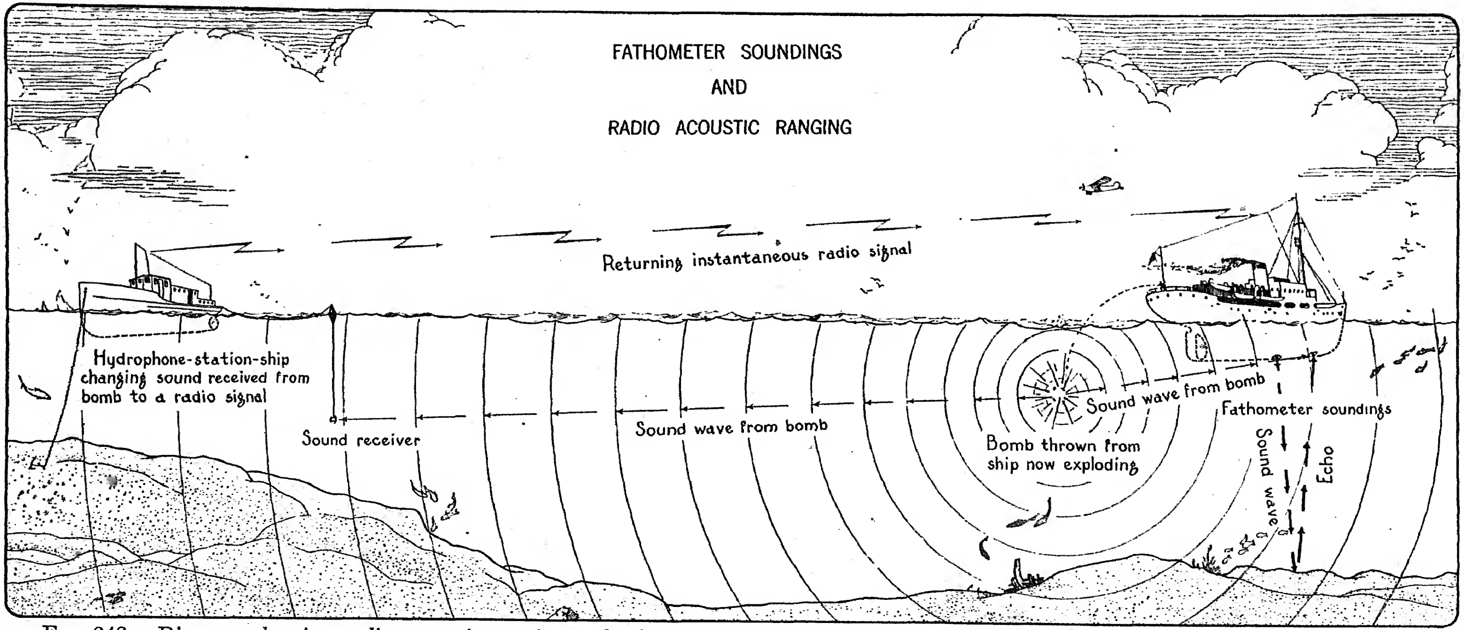

FIG. 242.—Diagram showing radio acoustic ranging and echo sounding as practiced in submarine topographic surveying. The time between the sound from the bomb as received on the survey ship and the radio wave from the station ship is measured on the survey ship to one one-hundredth of a second. Knowing the velocity of sound in water, the distance between the two can be calculated accurately, even if separated by as much as 100 miles. (*After U. S. Coast and Geodetic Survey.*)

Ocean Basin.—The depth of the ocean is determined by sounding. The unit of measurement is the fathom (6 feet), and the instrument employed in making the measurements is the *fathometer* (Fig. 242). It measures the time required for a sound to travel to the ocean bottom and the echo to return to the boat. Based on the velocity of sound waves in water the instrument automatically registers the depth as the ship steams along on its course. Such soundings are made at the rate of four per second, which is in marked contrast to the old laborious method with a weight and line dropped to the bottom. By the latter method the *Challenger* Expedition made only 504 deep-sea soundings during its cruise from 1873 to 1876. The locations of many of these old soundings have been found to be in error by as much as five miles, whereas by means of modern radio-acoustic ranging the soundings now being made are located about as accurately as are points on the average topographic survey of the land. The topography of the ocean bottom is destined to become almost as well known as is the topography of the land.

Although the depressed oceanic area within which the ocean lies is called a basin, it is noteworthy that, like the surface of the sea itself, the bottom is the surface of an oblate spheroid and is convex outward (Fig. 240). Water lies within it, because the points in the bottom are nearer the center of the earth than the adjacent points on the land. The ocean basin occupies about two-thirds of the earth's surface, but there is an excess of sea water, and the basin is somewhat more than full. Its waters, therefore, spread out over the low borders of the continental segments, covering them to depths which vary from near zero to 600 feet. At places this inundated continental fringe or border is more than 100 miles wide, and the total shallow-water portion amounts to 10,000,000 square miles. The whole ocean, including the basins and the shallow areas, covers 72 per cent of the earth's surface.

The ocean has an average depth of about 2½ miles, but half of it has a depth of 3 miles or more, and about 4 per cent of the sea bottom descends to the great depths which range from nearly 4 to more than 6 miles. The greatest known depths lie in the western part of the Pacific Ocean, where several soundings have been made that exceed 30,000 feet. The greatest depth was found just east of the Philippine Islands, where sounding shows more than 6 miles (35,433 feet) of water. The boundaries of these depressed areas at places are sharp, and the changes in the sea bottom precipitous. They are known as *deeps*, and some of them are probably due to block faulting. The ocean bottom as measured from sea level to the bottoms of the deeps shows greater relief than the land areas, but, in general, sedimentation tends to obliterate differences in level, and much of the ocean bottom is probably characterized by subdued topography.

The shallow portion of the ocean adjacent to the land is known as the *continental shelf* (Figs. 243, 244). It slopes gradually from the shore to about the 100-fathom line, where the bottom begins to descend more abruptly into abysmal depths. The continental shelf varies greatly in width along different parts of its extent and is continually being built up and out by the land-derived débris discharged by streams, by the wind, or by that worn from the coast line by the sea itself. Sinking of the coast line may cause the sea to spread farther over the low coastal areas of the continents, and for a time the shallow portion of the adjacent ocean may show features due to erosion. Thus there are channels extending out across the continental shelf while here and there terraces

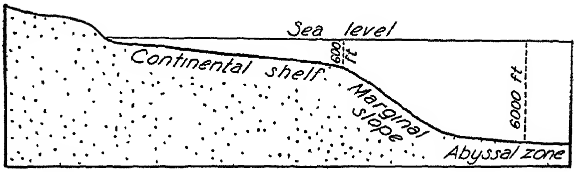

FIG. 243.—Diagram illustrating the relation of the marginal part of the ocean basin to the land.

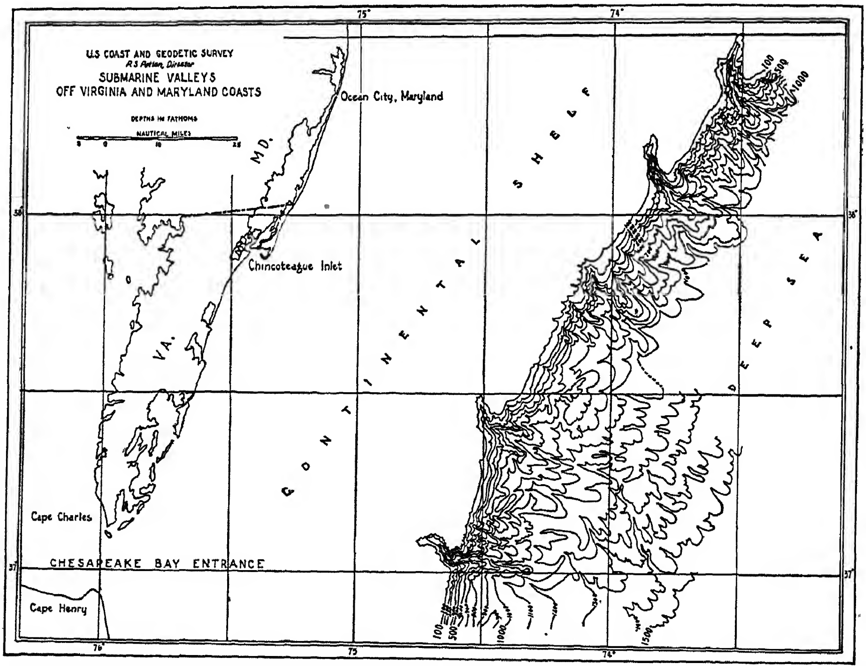

FIG. 244.—Submarine valleys off Virginia and Maryland coasts. (*After U. S. Coast and Geodetic Survey.*)

and ridges may be found. Some of these may have been cut by waves or built up by wave interference. Along the coast of New England the bottom in shallow water still retains many glacial features among which are drumlins, some projecting above water as islands.

The continental slope along those portions of North America where it has been accurately surveyed is not an even descent from the shallow water to abysmal depth but is a markedly rugged slope characterized by deep valleys with tributaries showing the common dendritic pattern such as that of stream erosion. It has been suggested that these deep channels and valley-like furrows leading seaward on the continental slope may have been produced by submarine mud flows or by the slumping of recently deposited highly water-soaked sediments on the steep outer depositional front when the angle of rest is exceeded. Certain of these gorges which have been under observation, such as Scripps Canyon off the California coast, occasionally show sudden deepening that seems to favor this explanation. Another suggestion is that "suspension currents," or currents of water heavily laden with finely divided sediment, may run down over the continental slope, below the reach of wave base, and produce a dendritic erosion pattern similar to that formed by streams on land. It is probable that both processes are active on the steep outer front of the continent.

At about the 100-fathom line the slope of the sea bottom, as stated, generally becomes more abrupt. The continental slope thus defined may vary in width from a few miles to 50 or more, but it is nearly everywhere a feature of the border of the ocean basin and is one of the most marked changes of level in the sea bottom.

The ocean bottom, beyond the continental slope, is still largely unknown. Here and there, however, ranges of mountains rise from the ocean floor, and the tops of these may form strings of islands. Thus Japan rises out of one of the deepest parts of the ocean, as do also the Philippine Islands. In midocean broad areas may rise thousands of feet above the general level of the surrounding sea bottom and extend ridge-like for great distances. An example is found in the middle part of the Atlantic, where such an area tends to parallel the eastern margin of North and South America.

Volcanoes erupt upon the sea bottom and build their cones above the level of the sea to form islands such as Hawaii. Most of the islands that dot the Pacific Ocean are partly of volcanic origin. The total number of submarine volcanoes is unknown, but 80 or more, the craters of which do not reach the surface, have been observed in action.

It is thus evident that the geologic processes affecting the surface of the continents—gradation, diastrophism, and vulcanism—are at work also on the ocean bottom.

Sea Level.—The surface of the ocean is the datum or reference plane for all topographic and geologic work. All level surfaces are theoretically parallel to mean sea level, and that surface is assumed to be constant. It is, of course, not level but is nearly the surface of an oblate spheroid. It is not exactly the surface of an oblate spheroid, because the mobile waters are attracted by, and drawn up against, the sides of the basin nearest the greatest land masses, such as the huge mountain mass of the Southwestern United States. Temporary changes in level, such as those produced by the tides, winds, and excessive rainfall, leave no permanent effect on the sea level but may produce marked effects on the coast lines where the waves reach above their ordinary height and carve out the rocks in advance of the normal sequence of events.

Composition of Ocean Water.—Sea water carries in solution great quantities of mineral matter, and thus it differs greatly from ordinary land waters. One thousand parts of sea water contain 34.4 parts by weight of mineral matter, or 3.44 per cent. For a cubic mile of sea water this amounts to 151,025,000 tons. Murray estimates the total amount of sea water to be 323,722,150 cubic miles; and since each of these cubic miles contains 3.44 per cent of salts, the total in the sea is over 4,500,000 cubic miles of mineral matter with a specific gravity of 2.2. This is equivalent to about 20 per cent of the volume of all rock masses above sea level and if precipitated on the sea bottom would make a layer about 175 feet thick over the entire ocean floor.

Salts Present in the Ocean
(After Dittmar)

	Per Cent
Chloride of sodium, NaCl	77.758
Chloride of magnesium, $MgCl_2$	10 878
Sulphate of magnesium, $MgSO_4$	4 737
Sulphate of calcium, $CaSO_4$	3 600
Sulphate of potassium, K_2SO_4	2 465
Carbonate of calcium, $CaCO_3$	0 345
Bromide of magnesium, $MgBr_2$	0.217
Total	100 000

About 32 elements have been found in sea water. In the order of amounts present in the sea some of these are chlorine, sodium, magnesium, oxygen, sulphur, calcium, and potassium. When the sea was originally formed, its waters may have contained some mineral matter in solution, and some may have been dissolved from the rocks of the area over which it spread, but probably the greater part of the solid matter has been contributed by streams, which obtained it by solution from the rocks of the land.

AVERAGE COMPOSITION OF SALTS IN RIVER AND SEA WATERS
(After Clarke, U. S. Geol. Survey *Bull.* 770)

Constituent	Percentage	
	River water	Sea water
Calcium	20.39	1.19
Silica, SiO_2	11.67	0.00
Sodium	5.79	30.59
Magnesium	3.41	3.72
Ferric and aluminum oxides	2.75	0.00
Potassium	2.12	1.11
CO_3 radical	35.15	0.21
SO_4 radical	12.14	7.70
Cl radical	5.69	55.48
NO_3 radical	0.90	0.00
Total	100.01	100.00

Mineral matter is constantly being removed from the sea water by marine animals that use it in forming their shells which are chiefly composed of calcium compounds. Still other parts of it have been removed by precipitation, either where the temperature has been rising and the carbon dioxide has been escaping or where excessive evaporation has been taking place. Evaporation causes also the formation of salt and gypsum deposits but probably is confined to shallow bays, coastal lagoons, and closed basins. Notwithstanding these removals, probably the mineral content of the sea is increasing. Common salt, NaCl, is accumulating in the ocean more rapidly than the calcium salts, because it is rejected by organisms in the making of their skeletons or shells and is removed by evaporation only where unusual conditions prevail.

In addition to the solids, gases also are held in solution by sea water. Among the more abundant ones are nitrogen, oxygen, and carbon dioxide. As an evidence of the amounts of these gases, one may consider carbon dioxide. According to T. Schloesing, the sea contains eighteen to twenty-seven times as much as is contained in the atmosphere. Assuming that air contains 3 parts of CO_2 per 10,000, it is estimated that the atmosphere contains 2,200,000,000,000 tons of carbon dioxide. The carbon in the carbon dioxide in the air is equivalent to about one and one-half times that contained in the estimated coal reserves of the world. The sea contains eighteen times as much as the atmosphere. The gases in sea water are chiefly gases that have been absorbed from the atmosphere, although some probably have been derived from submarine volcanoes and some from disintegration of organic matter or have been liberated from compounds through life processes. Oxygen is essential to nearly all life of the sea and to the process of oxidation in putrefaction, while carbon dioxide is the chief food material of the green and brown algae in the sea. All marine organisms use up and therefore tend to decrease the quantity of gas in solution in sea water. Cold water is capable of holding a greater quantity of gas than warm water. The gases absorbed from the atmosphere in the cold regions are diffused through the deep waters of the ocean and released in the warm regions of the earth or wherever these cold waters come to the surface and are warmed.

Temperature.—The temperature of the ocean is dependent almost entirely on solar radiation. The slight heat received from the earth's interior and from the radioactivity of the sea bottom is negligible in comparison. The temperature varies greatly from equator to poles and from surface to the abysmal depths. At the equator the surface temperature averages about 80°F., whereas in the polar regions it is about 28°F., or near the freezing point of sea water. The water at the bottom of the deep sea varies from the 28°F. of the polar regions to about 35°F. in the lower latitudes. According to Johnstone, the temperature of sea water in general changes rapidly from the surface to the 600-fathom line, where it is about 39°F., then less rapidly to the 1,000-fathom line, and below that it is relatively constant at or slightly below 35°F. The great body of the ocean is therefore cold, and the heavy cold waters originating in polar and subpolar regions, creeping equatorward, dominate the circulation of the modern sea and the general climate of the bordering lands.

Movements.—Water is constantly in motion in the "restless" sea (Fig. 245). Disturbances in any part of the ocean are transmitted through water to distant parts, and hence the ceaseless motion. One of the common causes is the drag or friction of the wind as it passes along the surface. Such movements are subject to marked changes, and they vary according to the strength and direction of the wind. During heavy storms these movements are awe inspiring, and their destructive power is very great. The pressure exerted by such waves may exceed a ton per square foot, and the height to which they may cause damage is illustrated by the lighthouse at Dunnet Head on the north coast of Scotland, where windows 300 feet above sea level have been broken by stones hurled up by the waves. Very severe storms may cause the sea to rise and move forward. Thus on October 5, 1864, a violent storm changed the level of the sea at Calcutta about 24 feet and inundated a large area. On September 8, 1900, a similar rise of the sea occurred during the Galveston storm; and also in Puerto Rico on September 13, 1928. Such storms are of the typhoon or hurricane type and may cause marked destruction along coasts. In 1935, a hurricane struck the keys off the south coast of Florida. The ocean swept through between islands where railroad embankments had been built, and the twisted rails with ties attached were left a hundred yards from their former locations. Much of the archway of the great inter-island viaduct was ruined, and both railroad and highway were washed out for miles. Waves rolled over parts of the land, and nearly completed the destruction. Buildings and steel towers fell before their attack, and in a well-populated section, where the storm was more severe, a lone house remained after it had subsided.

Earthquakes may cause waves of destructive violence. Such waves have been known to sweep inland 7 or 8 miles. Waves of this sort are usually referred to as tidal waves, although they have no connection with the true tides. The better term is sea waves.

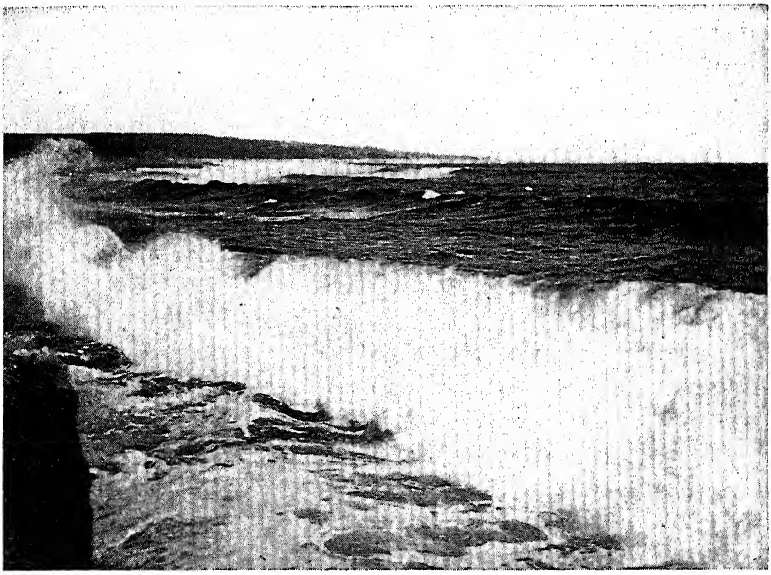

Fig. 245.—The surf along the Atlantic Coast of Nova Scotia. (*Courtesy National Development Bureau of Canada.*)

Rivers entering the ocean discharge great quantities of water at the coast line, where it tends to pile up. This comparatively warm, fresh water is lighter than the salt water of the sea, and it floats for a time as it spreads out and mingles with the salt water of the sea. The Mississippi and other rivers entering the Gulf of Mexico contribute largely to the Gulf Stream as it passes out into the Atlantic Ocean. Excessive rainfall on any part of the sea causes the water temporarily to pile up, and during its distribution motion is inevitable.

Evaporation removes great quantities of water, increasing the salinity of the surface waters and therefore the density of the sea. This is most pronounced at the equator. Along the coast of India it has been found that evaporation from the free surface of the sea amounts to about 23 feet per year. The cold, dense waters from the polar regions, which have crept equatorward and now fill the deep sea, rise and take the place of the vast quantities removed by evaporation, especially in the warm latitudes. Changes in density of sea water whether produced by evaporation, by the removal of calcium carbonate from sea water by lime-secreting life, or by other causes will produce motion, and all such changes contribute to the circulation of sea water. The attraction of the sun and moon causes a movement which is known as the tide.

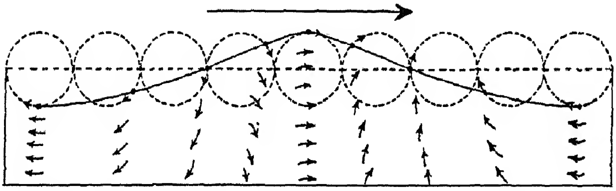

Fig. 246.—Diagram showing the motion of water particles in an oscillatory wave. The large arrow shows the direction of wave advance, the small arrows the directions in which water particles are moving in different parts of the wave. (*After D. W. Johnson.*)

Waves and Breakers.—The winds and other agents produce waves (Fig. 245). Wave motion is oscillatory. That is, each particle in the medium affected describes a circular orbit, returning approximately to the point of origin of the motion (Fig. 246). The diameter of the circle is determined by the height of the wave or the vertical distance between crest and trough. The length of the wave is the distance from crest to crest. The ratio of the length to the height determines the amount of motion that is transmitted downward. The motion travels downward to a depth undetermined; it decreases rapidly with depth and is generally imperceptible below 100 fathoms. The great storm waves travel out from under the region of disturbed conditions, and, although they preserve their length and velocity, they diminish in height as they pass into the ground swells of the sea. Waves approaching a

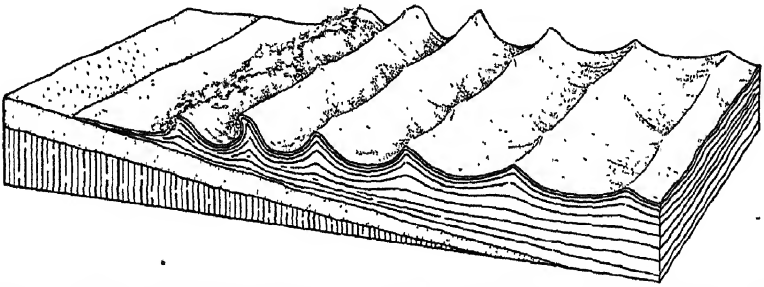

Fig. 247.—Diagram illustrating the drag, crowding, and breaking of a wave on a sloping or shelving coast.

shelving coast line drag bottom, and the lower part is retarded while the surface portion tends to maintain the velocity of the wave motion. The crest then becomes higher and the trough deeper until finally the crest moves ahead of its supporting column of water and breaks or tumbles down over the side of the wave into the trough, displaying a foaming surface (Fig. 247). Waves of similar height break in about the same depth of water, and thus a line of breakers is formed.

Local Currents.—After the wave breaks, the water rushes in upon the shore, and the remaining force is spent on the coast line. The smaller waves may strike without breaking and spend their entire force against the land. The water returning to seaward along the bottom and under the next incoming wave is known as the *undertow*. When waves strike the shore line obliquely (Fig. 248), two currents result: one at right angles to the coast line, producing the undertow, and the other parallel to the coast line, producing the *shore current*. Various modifications of these result from local conditions. Thus the undertow may not return normal to the coast line but may be deflected slightly in the direction of the shore current.

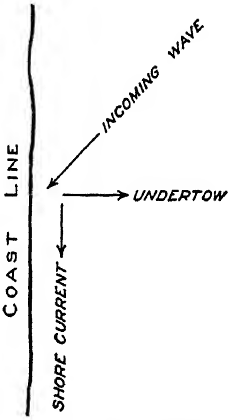

FIG. 248.—Diagram illustrating the formation of the shore or littoral current and the undertow.

Ocean Currents.—In regions where the winds have more or less constant direction, such as in the trade-wind belts, the surface waters are dragged along in the same direction with a velocity less than that of the wind itself (Figs. 249, 250). Since these winds blow from the northeast north of the equator and from the southeast south of the equator, the water is being urged toward the equator from both sides and there drifts westward as one current which would encircle the earth if the ocean covered the entire earth. This *equatorial drift*, contributed to and modified by numerous other factors, is the origin of the currents in equatorial oceans. Thus in the North Atlantic the westward-drifting equatorial waters strike the eastward-projecting portion of Brazil, and part is deflected up along the north coast of South America, striking the eastern terminus of the Greater and Lesser Antilles, where it is again divided, part of it crossing the Caribbean Sea and entering the Gulf of Mexico. Augmented by the great quantity of water being poured into the Gulf by rain and rivers, it emerges through the Straits of Florida as the Gulf Stream. Similar currents are formed in the South Atlantic; and in the Pacific Ocean, where the equatorial drift is more pronounced.

These ocean currents are not due to wind alone. The surface heating of the equatorial waters and their consequent expansion, and cooling in the polar regions, are in themselves sufficient to produce circulation. In the case of the Gulf Stream undoubtedly the large rivers pouring into the gulf are important. Far out at sea that stream can be recognized by its color, due in part to the excess of fresh water which has been poured into it. The shape of the continental shelf, the configuration of the coast line, and the rotation of the earth modify these currents and control their movements to some extent; but as long as the generating forces act, they keep moving. The chief effects of such currents are climatic, as they carry little sediment.

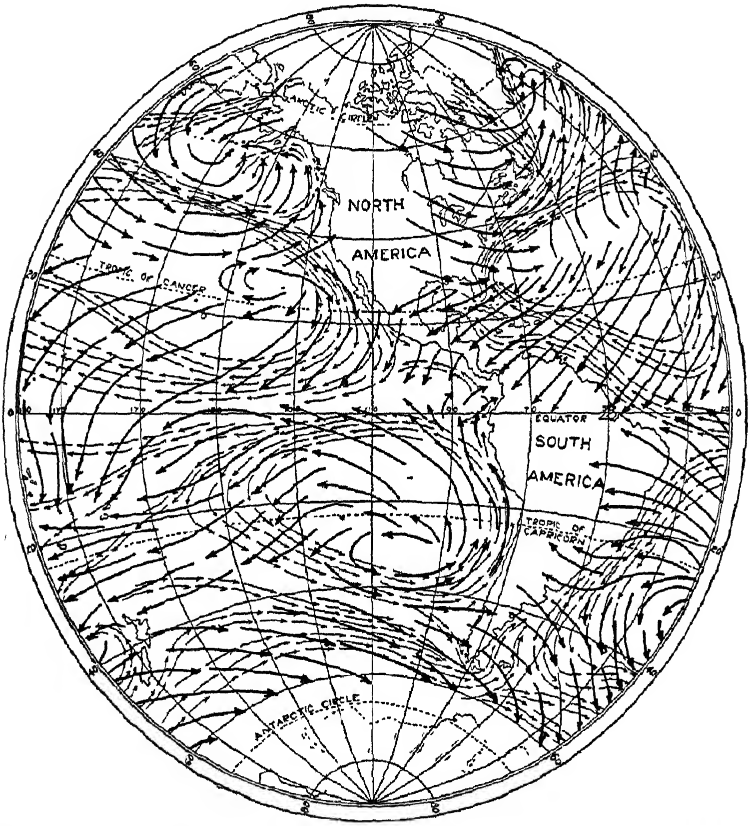

FIG. 249.—Map of the western hemisphere showing the prevailing winds (large arrows) and the ocean currents (small arrows) during the month of January. (*Modified from W. M. Davis.*)

Tides.—The periodic rise and fall of the sea, rising twice in 24 hours and 52 minutes, is the tide (Fig. 251). It is produced by the differential attraction of the sun and of the moon on the earth. This attraction is effective on both land and sea, but the sea is more mobile and therefore yields more readily to the pull thus set up.

The sun, moon, and earth lie in nearly the same plane, but they continually shift their positions with reference to each other. The total gravitative effect of the sun and moon on the earth is constantly acting, but the pull is in different directions as their positions change; hence the effect on the earth varies from time to time.

Because of its nearness the moon produces the most marked results. The sun's tide acts more as a modifier of the moon's tide, and when these two bodies are in the same straight line, the effects are added and the result is an unusually high tide. Considering only the more effective force, the surface of the sea is pulled out or caused to bulge on the side nearest the moon, while on the opposite side of the earth a similar, but slightly smaller, bulge is produced.

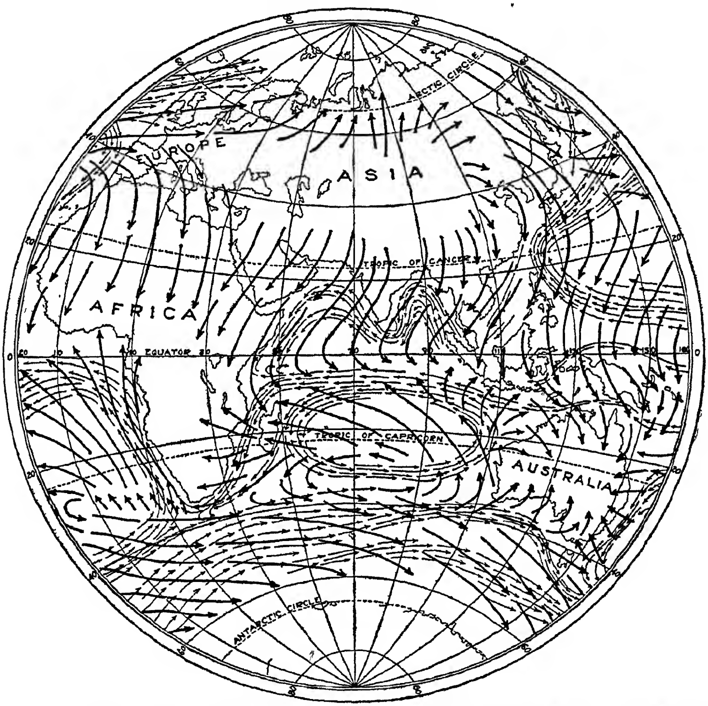

FIG. 250.—Map of the eastern hemisphere showing the prevailing winds (large arrows) and the ocean currents (small arrows) during the month of January. (*Modified from W. M. Davis.*)

In the open ocean the tidal variation of sea level probably does not exceed 2½ or 3 feet, but along the continental borders the differences between high and low tide are much greater, whereas in narrow bays with broad seaward openings, such as the Bay of Fundy, it may exceed 20 or even 50 feet. The tide usually comes in as a series of waves each reaching higher and higher until the crest of the rise is attained, after which recession sets in, and this continues until low tide. At certain places, however, the rise of the tide is sudden, coming in as a wall-like wave of water, known as a *bore* (Fig. 252), which may be 25 feet high. The bore is especially well developed on the coasts of China and India. Wherever the tide runs through restricted passages, such as narrow straits or bays or between islands, currents are developed which scour the bottom and shift about the loose materials. The tide is not a very important agent of erosion. There are broad belts of the surface, adjacent to the sea, however, that are usually referred to as tidal flats. These are

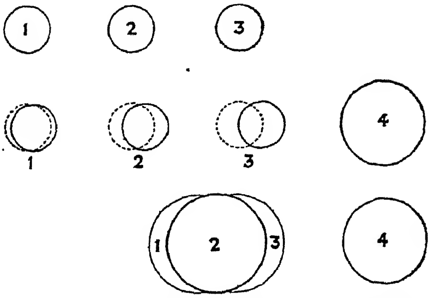

Fig. 251.—Diagrams illustrating the origin of tides. Suppose the bodies 1, 2, 3 are affected by the presence of 4; 1 is pulled toward 4 a short distance, 2 is pulled toward 4 a greater distance, and 3 is pulled toward 4 a still greater distance. In the lower figure if the water 1 on the far side of the globe is pulled toward 4 a short distance and the globe 2 is pulled a greater distance, the globe 2 is pulled away from the water 1. The water at 3, being nearer the attracting mass, is pulled farther than 2 and is pulled away from 2. The earth 2 is pulled away from the water 1, and the water 3 is pulled away from the earth 2.

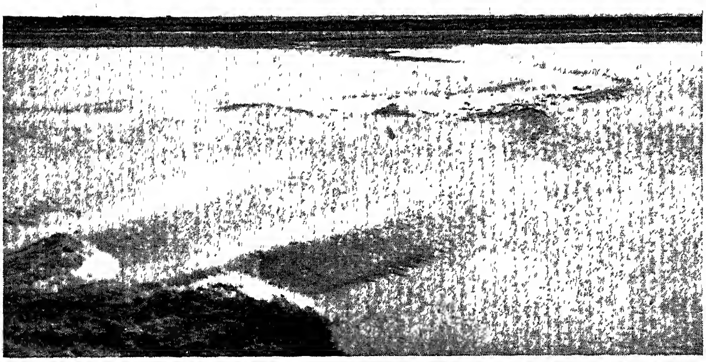

Fig. 252.—The tidal bore passing Moncton, New Brunswick. (*Courtesy Can. National Development Bureau.*)

covered by water part of the time and during the remainder of the day or night are exposed to the ordinary weathering processes (Fig. 253).

Fig. 253.—A submerged forest and tidal flat at Fort Lawrence, New Brunswick, forming a part of the littoral zone. When the tide is in, these stumps are covered by 25 feet or more of sea water. (*Courtesy Can. Geol. Survey.*)

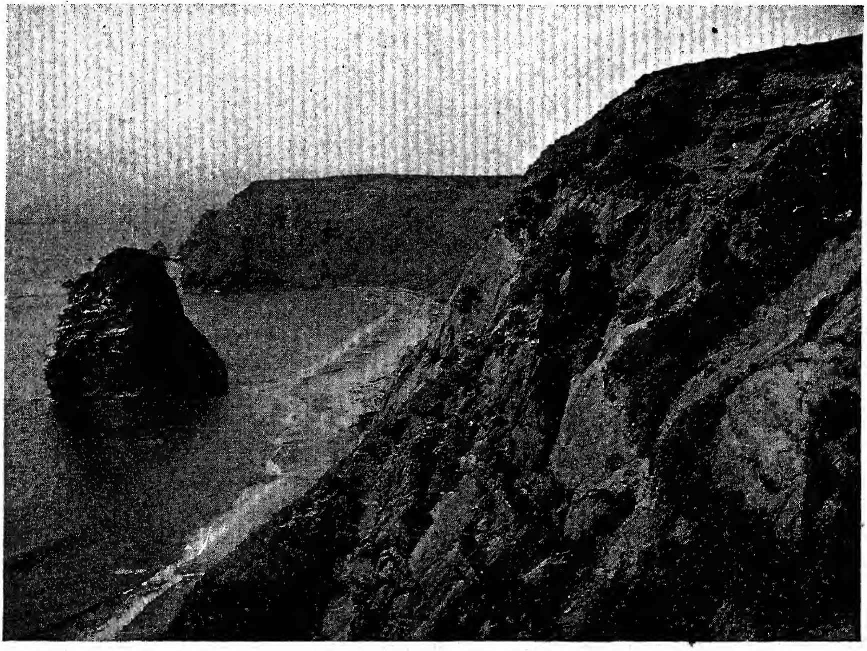

Fig. 254.—The sea cliff at Cape Blanco, Oregon.

Mechanical Effects of Moving Sea Water.—Movements in sea water may produce mechanical effects of vast importance owing to the mass of the moving water and its velocity. Thus the waves beating upon a coast gradually wear it down. This is accomplished largely by grinding. The rock fragments quarried out by the waves or rolled down into the water are hurled back by the waves against the coast line and are thus effective tools in cutting the shore line or undercutting promontories.

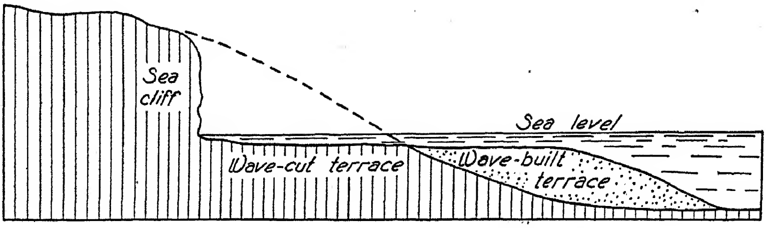

Fig. 255.—Diagram illustrating the development of a sea cliff, a wave-cut terrace, and a wave-built terrace. The portion of land below the dotted line has been removed by wave erosion.

The overhanging rock then topples into the sea, and more tools are supplied to continue the work. The vertical cliff thus produced is known as a *sea cliff* (Fig. 254). Such cliffs, on the south shore of Nantucket Island, have been cut back by the waves as much as 6 feet per year. That portion of the land over which the sea has advanced by cutting away the cliff is known as the *wave-cut terrace,* while the continuation of this surface to seaward produced by the deposition of the materials, resulting from the destruction of the cliff, is known as the *wave-built terrace* (Figs. 255, 256).

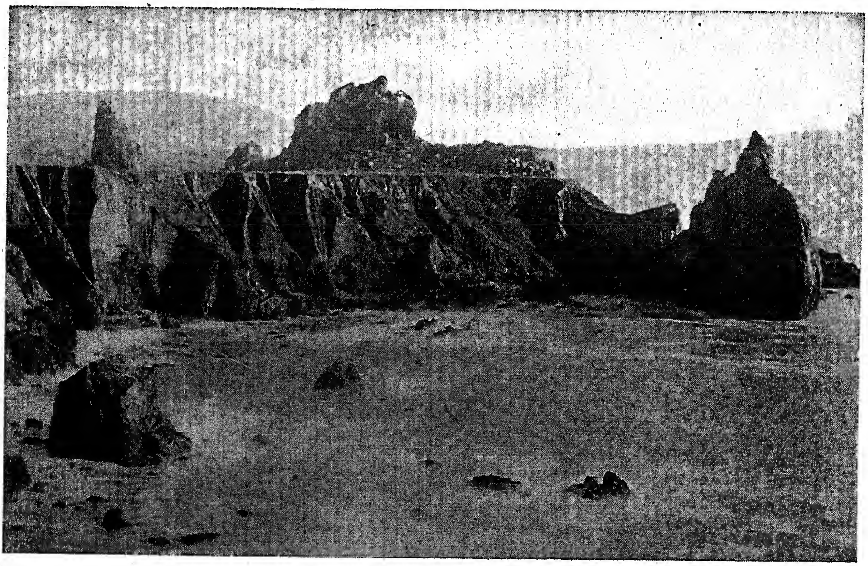

Fig. 256.—Sea cliff and elevated wave-cut terrace north of Port Harford, California. (*Photograph by Stose, U. S. Geol. Survey.*)

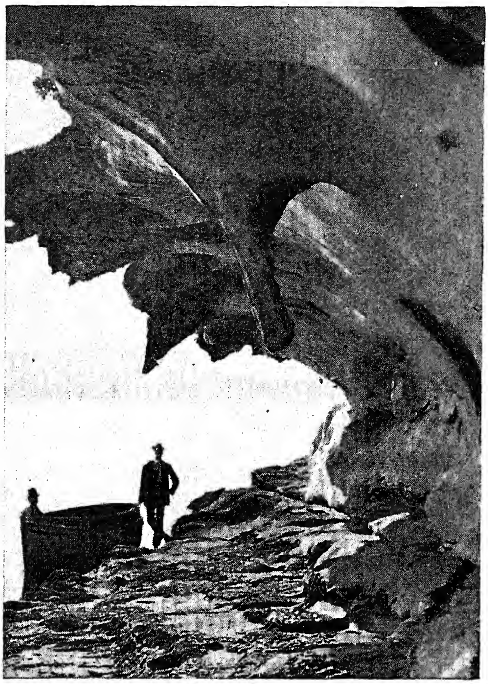

FIG. 257.—The wave-notched base of a sea cliff near Nanaimo, British Columbia. (*Courtesy Can. Department of Interior.*)

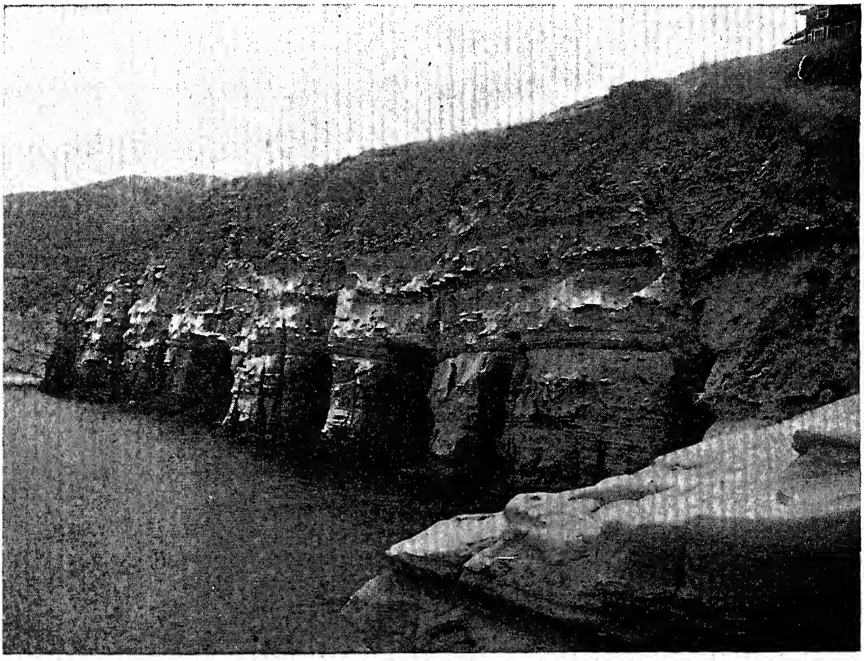

FIG. 258.—Sea cliff and caves at LaJolla, California. (*Photograph by Arnold, U. S. Geol. Survey.*)

Sea Caves.—Incoming waves which strike a bold coast line fill crevices in the rocks and force them apart by hydraulic pressure that may exceed a ton per square foot. Particles of the rock are removed, blocks are quarried out, and thus *sea caves* are formed (Figs. 257, 258).

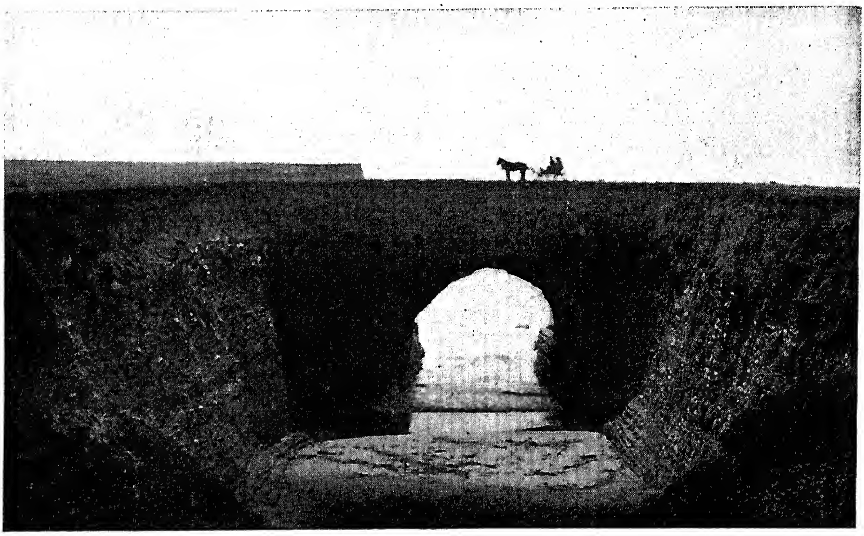

Fig. 259.—Natural bridge, near Santa Cruz, California, formed by collapse of the roof in a sea cave. (*Photograph by Arnold, U. S. Geol. Survey.*)

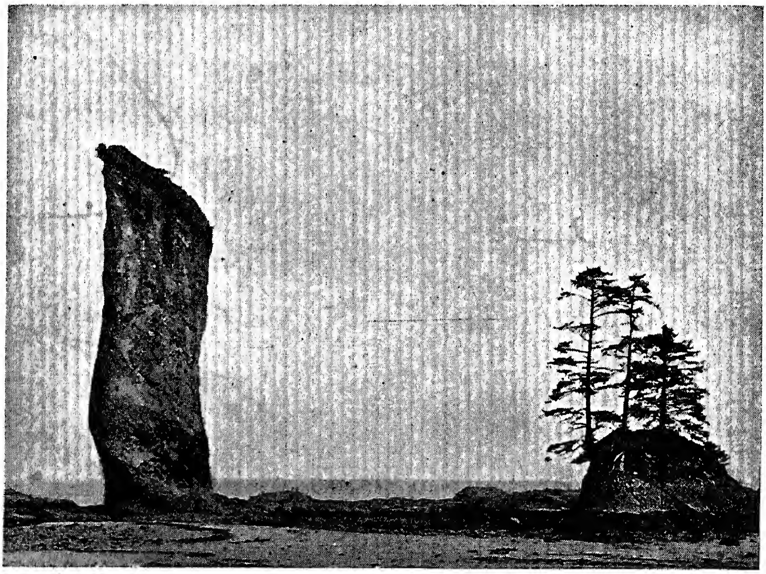

Fig. 260.—A stack at Graham Island, British Columbia. (*Courtesy Can. Geol. Survey.*)

In such cavities the alternate compression and rarefaction of the air, with the advance and retreat of the wave, may break through the roof

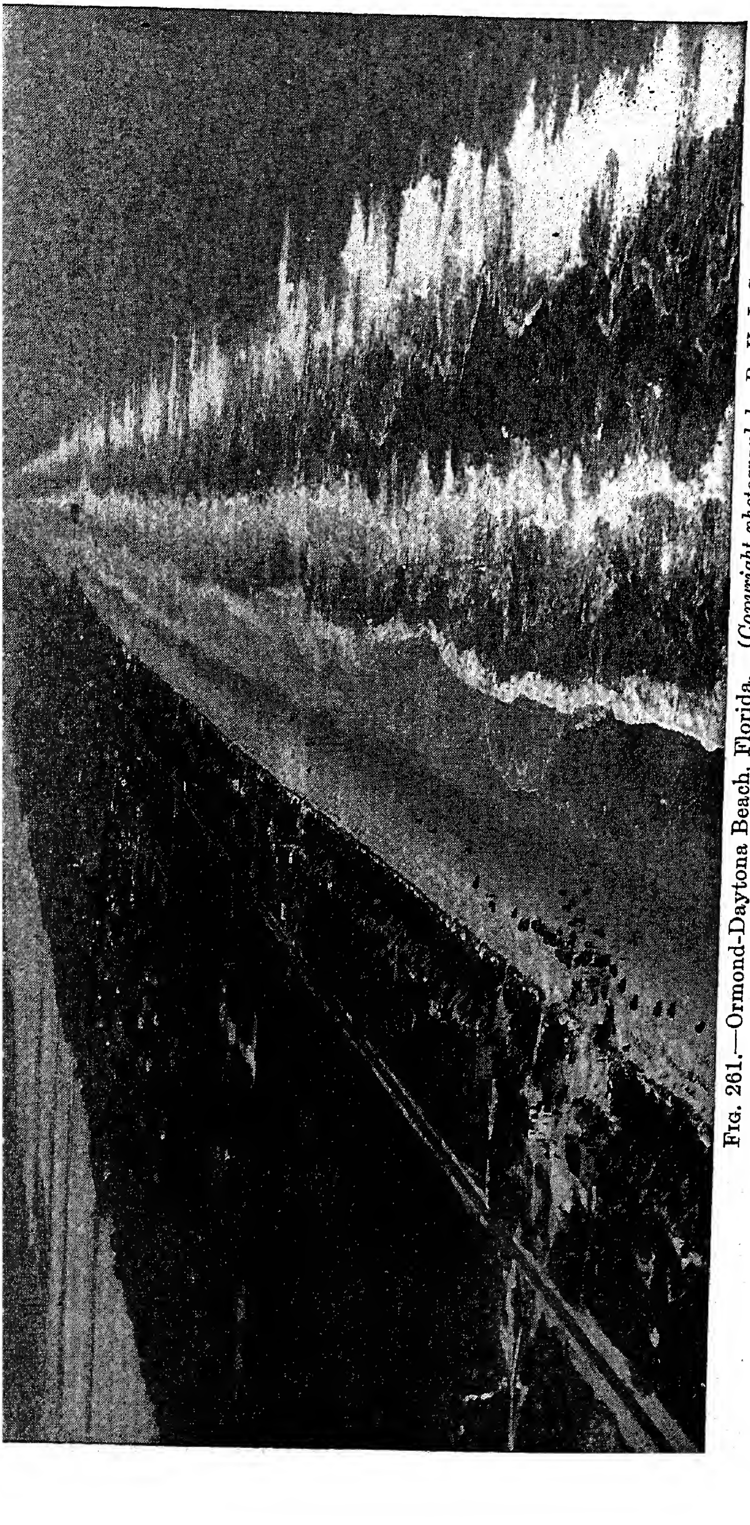

Fig. 261.—Ormond-Daytona Beach, Florida. (*Copyright photograph by R. H. LeSesne.*)

or top of the cave and form a *spouting horn.* Sea caves are found commonly along the New England coast but may be expected on any rocky shore line subjected to vigorous attack by waves. Occasionally caves formed on opposite sides of a narrow promontory may break through, forming a pierced rock or bridge (Fig. 259). When the bridge or arch falls, the seaward mass becomes an island, usually known as a *stack* (Fig. 260). On rocky coasts the returning water from the spray and splash of storm waves rotates the water and sediment caught in small

Fig. 262.—Shingle beach of Lake Ontario near Olcott, New York. (*Photograph by Gilbert, U. S. Geol. Survey.*)

Fig. 263.—The upper margin of a wave leaving its mark on the sand beach and the water returning as undertow at Balboa, California.

rock basins, grinding them deeper and producing the familiar tidal pools. Jug-shaped holes of such origin are known on the north shore of Lake Superior, but they are more common on ocean coasts.

Beach.—The beach (Fig. 261) is the area along the sea extending from the line reached by high tide and the highest storm waves to the low-tide mark. Twice in a little more than 24 hours most of it is swept by the tide and in the interim exposed to the heat of the sun or subjected to cooling at night. Shells and rocky materials are dragged up and down by the moving water and thus ground finer and finer. Occasionally when swept by violent storm waves the fine materials are removed from the beach and deposited far out from the shore line. The region is inhabited by both land and marine life. Thus the remains of both mingle and may be preserved as the typical fossil record of the time. Sediments also may mingle. The wind-blown sands of the dunes grade into those of typical marine origin, and river sediments may do likewise. Eventually the deposit, accumulating under such conditions, will bear all the marks of shallow-water deposition together with some more characteristic of land formations. Where the beach is composed largely of flattened, disk-like pebbles produced by a gliding wave motion, these may overlap forming a *shingle beach* (Fig. 262). Certain beaches are composed of shell fragments or coral sand, for example the "pink beach" of Bermuda, or of olivine sand, as in the Bay of Naples, but by far the greater number are mainly quartz sand.

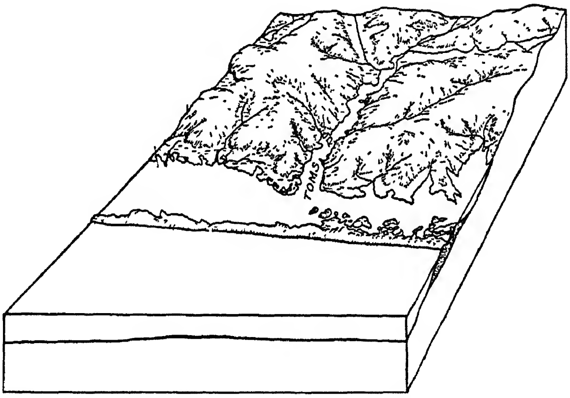

FIG. 264.—Block diagram of a barrier along the New Jersey coast, Asbury Park quadrangle.

Barriers, Bars, and Spits.—On a gently sloping coast where the base of the incoming wave begins to drag on the sea bottom it carries along the loose sediment found there and drops it where the wave drags and breaks. Thus to the breaker line the loose sediment on the bottom is carried from the seaward side, while on the landward side the returning water drags the loose materials to the same line (Fig. 265). A ridge is thus built up at the average breaker line which eventually the storm waves may build above sea level, and the wind may whip its sandy surface

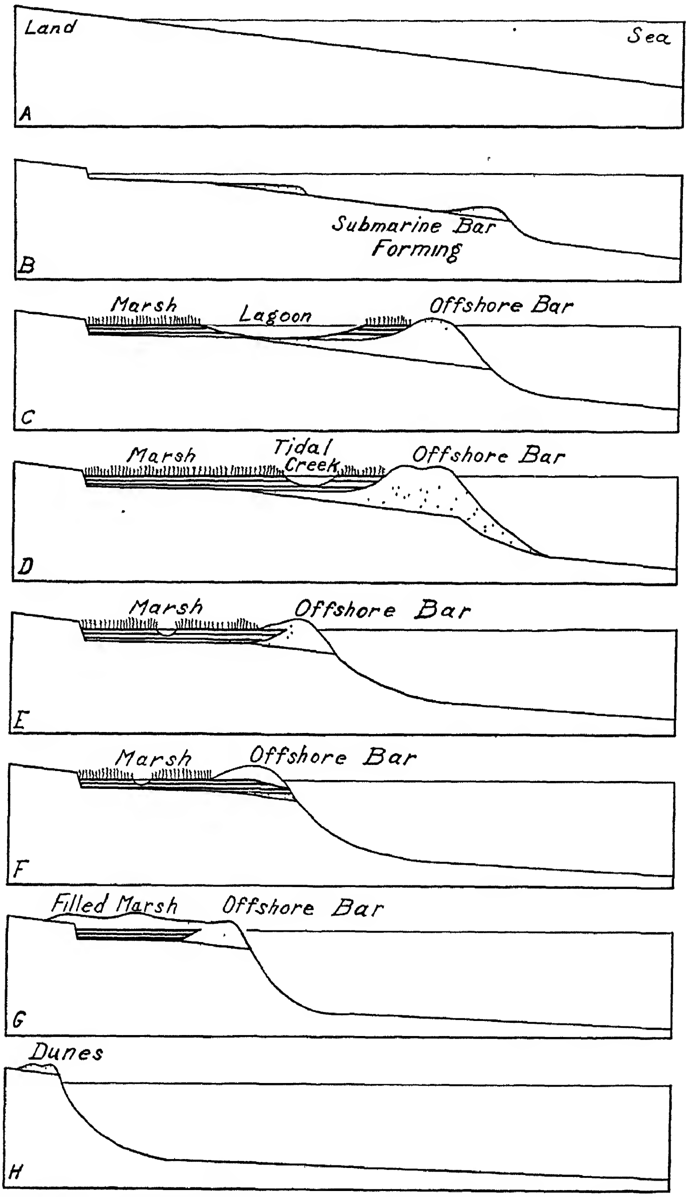

FIG. 265.—Diagrams illustrating the development of and later the destruction of an offshore bar and lagoon. (*Based on figures by D. W. Johnson.*)

into dunes. Such a ridge built above sea level at the old breaker line is known as a *barrier*, or *offshore bar* (Figs. 264, 265). The area behind the barrier is shut off from the open ocean. If it is gradually filled in by detritus carried into it from the land by streams or blown in by winds, it becomes a swampy tidal flat and finally, when it is completely filled, a part of the land. The coast line is then transferred to the old breaker line. This condition is common along the Atlantic Coast of North America from New Jersey southward.

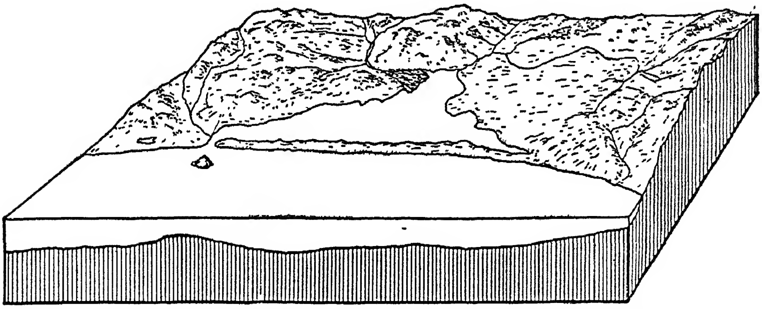

FIG. 266.—Diagram illustrating the bar built nearly across the entrance to the bay at Morro, California.

The shore current shifts sediments parallel to the coast line, but owing to the tendency of a current to continue in a straight line it fails to follow the indentation that may be produced by a drowned river valley or by a bay of any origin. Where the shore or littoral current passes from shallow to deeper water at the entrance to such indentations or bays, deposition is almost certain to take place. A ridge which becomes a land projection, or *spit*, is thus built up. When a spit is built almost or entirely across the entrance to the bay, it becomes a *bar* (Figs. 266,

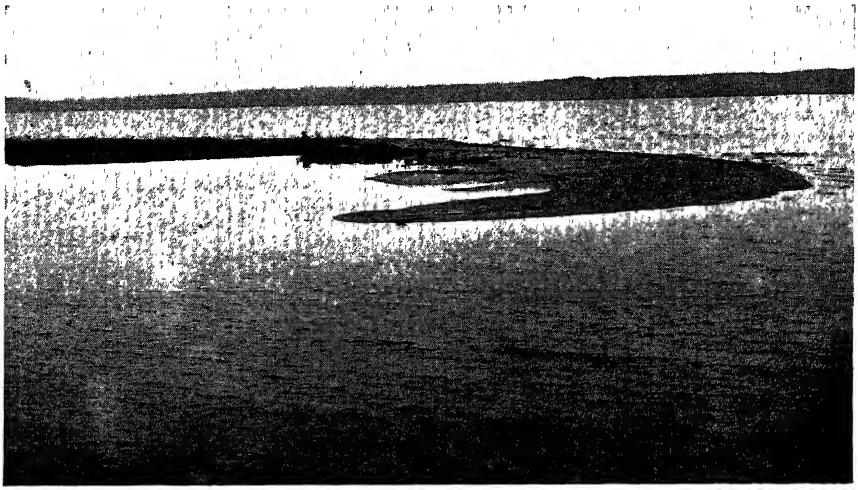

FIG. 267.—Recurved spit or hook, Duck Point, Grand Traverse Bay, Lake Michigan. Beyond the left end of the view the spit is attached to a point of land. (*Russell, U. S. Geol. Survey.*)

267). Bars cut off the indentations and tend to simplify the coast line. Islands become connected with the mainland in a similar manner. Such islands are said to be land tied, and the bars acting as the lines of connection are called *tombolos* (Fig. 268). They are numerous along the New England coast, where several islands may be tied together, and the string thus formed may be connected with the mainland. Nahant, Massachusetts, is a land-tied island. The Rock of Gibralter is similarly united to the coast of Spain.

If the free end of a spit is beaten by violent storm waves or by those of seasonal storms, it may be cut back, and the resulting materials may be deposited behind the spit to form a *hook*. By prolonged cutting and deposition the curved end of the hook may become extended until it reaches the mainland, thus forming a *loop*. All of these constructional features, formed by waves and currents are subject to change throughout the years; and a spit, bar, or barrier may show very different outlines within a decade.

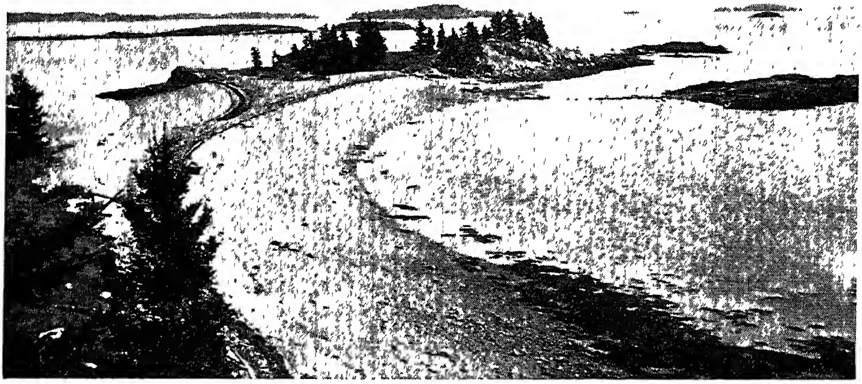

FIG. 268.—Land tied islands, Spruce Head Islands, Maine. (*Photograph by Bastin, U. S. Geol. Survey.*)

Shore Lines.—Shore lines record the history of the various geologic processes that played a part in shaping the outlines of the coast. Two distinct kinds of shore lines are evident. One is bold, irregular, and rocky like that of New England; and the other is smooth, low, and sandy like that of the southern Atlantic Coast of the United States. The first type is produced by partial submergence of the coast and is a *shore line of submergence.* The second results from a partially emerged sea floor and is a *shore line of emergence.* Other shore lines do not show the effects of either submergence of former land or emergence of a former sea floor. Such shores are classed as *neutral shore lines.* Still others show features that are a combination of at least two of the preceding classes. They are *compound shore lines.*

The erosional and depositional features of a shore line depend upon the amount of work accomplished by waves and currents. However formed, the coast may be in its initial, youthful, mature, or old stage of development. These terms are significant of the development toward stability of the coast line, and each stage has its own peculiar characteristics.

Shore Line of Submergence.—The initial stage of a shore line of submergence is characterized by deep bays, bold headlands, and islands. If the partially submerged land was well dissected by streams, or if it had been deeply eroded by glaciers, the rise of the sea floods or drowns the valleys, and they become embayments of the sea. During the youthful stage sea cliffs are cut on the exposed headlands and on the seaward shores of the islands. The resulting sediments are deposited as spits and bay-mouth bars in the deeper waters of the bays (Fig. 269). The larger dismembered tributaries of the drowned valleys build deltas in the sheltered portions of the bays. As these processes continue, the islands are cut away, and the headlands recede. Spits and bars simplify the outline of the shore line, and it approaches maturity. The mature stage of a shore line of submergence is characterized by high, straight sea cliffs and wide wave-cut terraces. The terraces extend seaward as far as the original headlands that were removed by wave erosion. Beyond these rocky platforms, wave-built terraces are deposited. The shortened bays are filled with delta sediments, and the shore line is shortened and simplified to the greatest extent. It is then mature. After this stage is reached the shore line retreats slowly, and the wave-cut terraces are

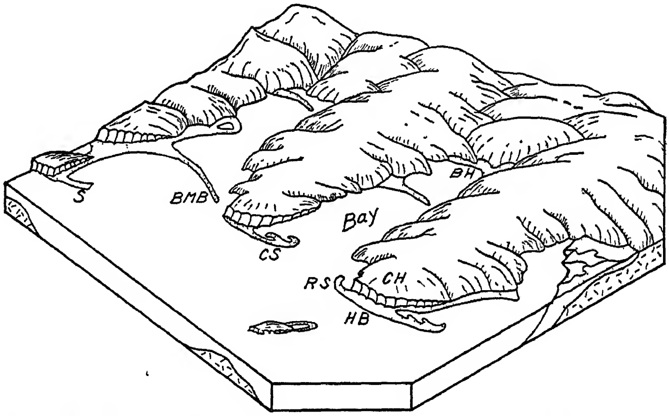

FIG. 269.—Diagram illustrating the land forms developed along a coastline of submergence. *S*, spit; *CS*, complex spit; *RS*, recurved spit; *HB*, headland beach; *BH*, bay-head beach; *CH*, cliffed headland; *BMB*, bay-mouth bar. (*After D. W. Johnson.*)

widened until the force of wave action is lost as the waves roll over the wide rocky platforms. Very little work is accomplished, and the shore has reached the stage of old age.

Shore Line of Emergence.—The initial stage of a shore line of emergence is a nearly flat, coastal plain covered with unconsolidated marine sediments. Most of the irregularities of the sea bottom had been obliterated by deposition of sediments. Erosion, however, begins immediately, and the stage of early youth is characterized by the development of barrier beaches or offshore bars along the line of breakers. The position of the bars is determined by the seaward slope of the sea floor and by the size of the waves. Shallow water lagoons lie between the offshore bars and the shore line (Fig. 270). The lagoons are gradually filled by sediments from the land, by the growth of vegetation, and by other less evident agencies. During the stage of late youth the waves continue to beat on the bars and to drive them landward. The lagoons are changed into tidal marshes (Fig. 265). As wave action continues, the marshes are filled completely, the offshore bars are eliminated, and the shore line is made straight and simple. It is then similar to the mature stage of a shore line of submergence. During the mature and old stages wave-cut terraces are developed as the shore moves landward.

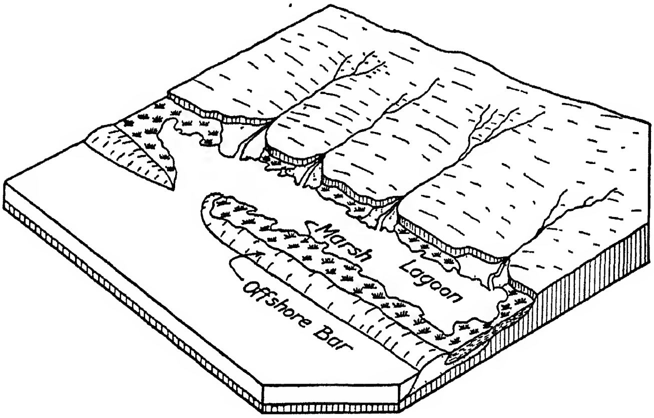

FIG. 270.—Diagram illustrating a coast line of emergence. (*After D. W. Johnson.*)

Neutral Shore Lines.—There are various classes of shore lines whose characteristics are not dependent on either submergence or emergence. These include (1) delta shore lines (Figs. 154, 227), (2) alluvial plain shore lines, (3) outwash plain shore lines, (4) volcano shore lines, (5) coral-reef shore lines, and (6) fault shore lines.

The shape of a delta shore line is determined by the currents that predominate. Where the currents of a river's distributaries are stronger than the littoral or shore currents, a lobate delta like that of the Mississippi River is formed. Where shore currents and wave erosion influence the deposition of delta sediments, a small delta modified by currents like the delta of the Tiber River results. Glaciers may leave moraines along the shore or in the shallow portion of the sea. Sand dunes on shore may shift and invade parts of the sea. Volcano shore lines result from the spread of lavas as is now taking place near Mauna Loa in Hawaii. Likewise the growth of volcanic cones along the coast changes the outlines of the shore. This is true especially of the highly explosive type of volcanoes such as many of those along the borders of the Aleutian Islands and on the South Sea Islands. Fault scarp coasts are characterized by the absence of an off-coast shallow area. The waves approach the coast from deep water and undercut the cliffs, producing wave-cut terraces. When the débris has partly filled the offshore area, wave-built terraces may form. Subsequent elevation of the land may rejuvenate the coastal area, raise the old cliffs, and cause new ones to be formed. Thus the shore line may reach maturity without developing any offshore bars or other features of the normal, gently sloping continental shelf.

Each class of neutral shore lines has characteristic features developed as it passes from youth to old age. Such shores, however, constitute but a small proportion of the shore lines of the earth.

Compound Shore Lines.—Some shore lines have complex histories. A cycle of development may be interrupted before it reaches maturity. This may be accomplished by either subsidence or uplift or by agents of gradation. The shore line of North Carolina records such interruption. It has the drowned valleys of a shore line of submergence and the offshore bars of a shore line of emergence. Such a combination of shore features indicates that extensive submergence was followed later by slight emergence which left the valleys submerged but reduced the depth of the water so that the waves dragged on the floor of the sea and developed offshore bars or barriers.

Life of the Sea.—It is usually supposed that the first life on the earth appeared in the sea or in the brackish waters of the coastal swamp. With its introduction came biochemical activity. Many of the organisms which live in the sea have hard parts made up of calcium carbonate or of silica, and these are the materials which form the limestones and siliceous deposits. It is largely through living organisms and their ability to precipitate calcium carbonate and amorphous silica that the cementing materials for other sediments are obtained. The rock-building life of the sea therefore is a geologic factor of great importance.

The common lime- and silica-secreting forms of life belong chiefly to the lower plants, such as the bacteria, the diatoms, and the algae, and to the invertebrates among the animals. It seldom happens that bacterial remains are long preserved in recognizable condition in the deposits which they are forming. Ancient limestones now wholly devoid of fossils may owe their origin to these organisms. Few of the marine plants, even of the algae, can be identified among the fossils so commonly occurring, and yet in the modern seas the calcareous algae can be found in the active process of rock building, often even exceeding the animal life of the region in the rapidity of their work. Thus on the typical coral island of Funafuti in the south Pacific two kinds of algae were found to have formed more of the rock than all of the corals combined.

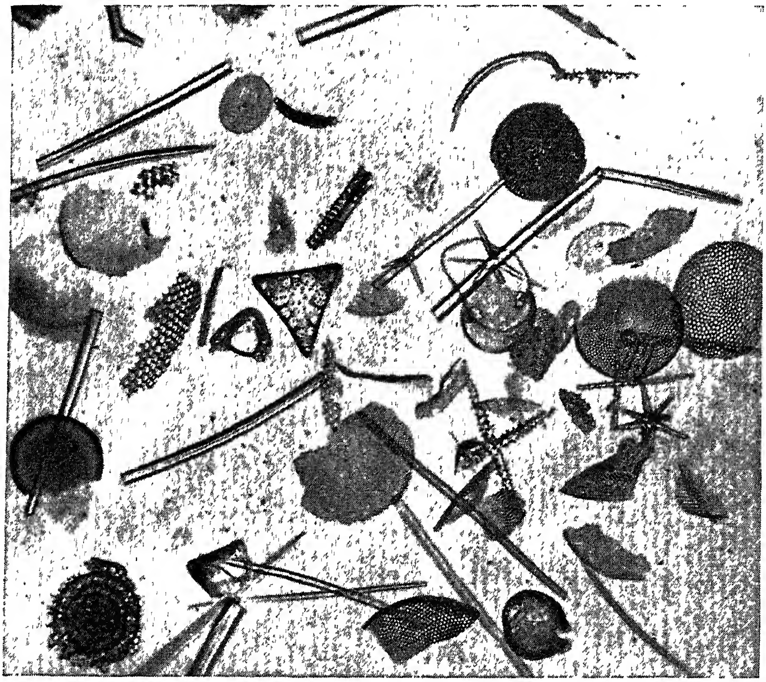

FIG. 271.—Diatoms and Radiolaria in recent ooze. (Greatly magnified.)

The diatoms have siliceous tests or capsule-like shells (Fig. 271). These accumulate in abundance under favorable conditions and give rise to thick deposits of "diatomaceous earth," or their siliceous material may be carried off in solution and contribute to the formation of flint or chert beds in adjacent limestones. The Radiolaria and certain of the sponges have siliceous skeletons which contribute to the formation of the same flints and cherts.

The lime-secreting animal life of the sea includes almost every group from the Foraminifera to the highest division of the invertebrates. Their remains are found in abundance in the deposits they form. The most prolific limestone builders occur among the Foraminifera, the Hydrozoa, the Anthozoa, the Crinoidea, the Echinoidea, the Bryozoa, the Brachiopoda, and the Mollusca in general. Other animal groups also make contributions but are seldom of such importance as those just mentioned. Certain areas of the shallow sea bottom, especially those comparatively free from land detritus, are crowded with these groups, and their remains accumulate in abundance to form limestone (Fig. 272).

Reefs.—Certain corals and other animals which live in colonies are reef building. A reef is a belt along which the corals live, their skeletons building up a platform on which the colony continues to grow. Reefs of this sort have been an important factor in the formation of limestone since very ancient (Ordovician) time and still are of common occurrence in the warm, shallow waters of the tropical and subtropical oceans. The total area covered by growing coral reefs is estimated to be 500,000 square miles, and the detritus derived from them by wave action may cover more than twice that area. Modern reef-building corals are confined to clear sea waters with a temperature that does not fall below 68°F. and a depth that varies from a little below mean sea level to about 150 feet. They are thus a shore phase of islands or of low shelving coasts of continents. The most vigorous growth of the reef is toward the open sea, where waves bring food, oxygen, and the necessary calcium for their skeletons. Opposite the mouths of rivers where quantities of fresh water and land detritus are discharged there is a break in the coral growth; tidal movements may preserve openings through the reef so that they form a disconnected belt along the coast. Massive reefs built along the shore are

FIG. 272.—Foraminifera in Cretaceous marl from New Jersey. (Greatly magnified.)

termed *fringing reefs.* Those that are separated from the shore by a channel or lagoon are called *barrier reefs.* Those that surround lagoons and are more or less circular in shape are called *atolls.* Darwin suggested that atolls are formed at certain stages in the history of a fringing reef surrounding an island on a sinking sea bottom (Fig. 273). The fringing reef, which grows chiefly to seaward, is changed to a barrier reef separated from land by a lagoon as the island sinks; and then as the island sinks farther and is submerged below sea level, the corals continue to build up the fringing reef, and it becomes an atoll. The waves break off fragments of the reef and pile them up above the sea level on the lagoonward side. In time, vegetation gets started, land is thus formed, and a ring-like coral island results. The most extensive coral reef of the present time is the great barrier reef along the northeast coast of Australia. It is more than 1,000 miles long and varies from 10 to 90 miles in width. Behind this barrier is a channel 20 to 50 miles wide and 60 to 240 feet deep, which is thus protected and is used for coastwise shipping, while the adjacent coast also is freed from the violence of the waves.

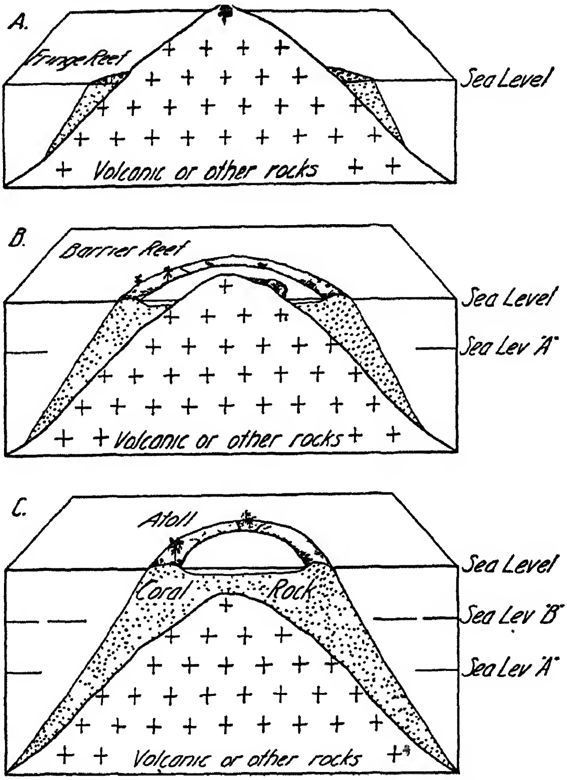

FIG. 273.—Diagrams showing how an atoll may be developed. *A*, island with a fringing coral reef; *B*, same after subsidence; the fringing reef has grown upward and become a barrier reef; *C*, same after further submergence of the original island; the barrier reef has become an atoll.

Ancient coral reefs are often found in the older limestone deposits. Some of these, as at Alpena, Michigan, and Formosa, Ontario, are lenticular and, because of their more rapid accumulation, develop minor bulges or upfolds in the succeeding strata. Other forms of life, such as lime-secreting Algae, Bryozoa, and Serpula, may build reefs, but the details of formation do not differ greatly from those operative in forming coral reefs.

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CHAPTER IX

THE ORIGIN AND NATURE OF SEDIMENTS

Origin of Sediments.—On the basis of origin, sediments may be classified as (1) land derived or terrigenous, (2) organic, (3) volcanic, and (4) magmatic. Of these the land-derived gravels, sands, and clays are by far the most abundant. They are formed by the disintegration and decomposition of former rocks and represent the end products of erosion.

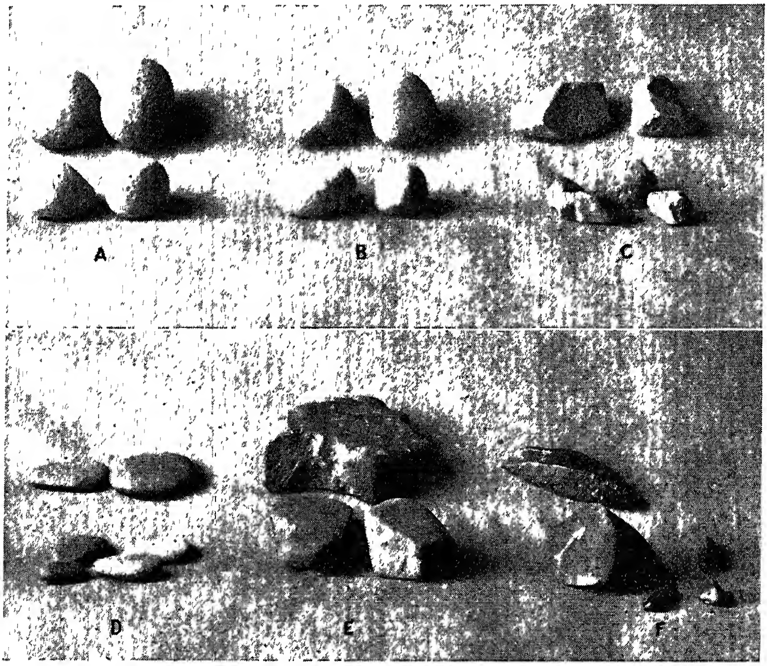

FIG. 274.—The shapes of pebbles. *A*, rounded; *B*, subrounded; *C*, angular; *D*, disk-like; *E*, faceted and striated glacial; *F*, dreikanter, faceted by windblown sand. (About one-fourth natural size.) (*After Grout.*)

Organic sediments are those formed from constituents that were once dissolved in water and later extracted through the activity of plants and animals. Many organisms use inorganic substances in the development of their protective and supporting structures, such as bones, shells, and tests. These structures contain phosphates, sulphides, iron oxides, lime and magnesium carbonates, silica, and other constituents in varying amounts that accumulate as sediments when the organisms perish. Other organisms bring about chemical reactions that lead to the precipitation of sediments.

Sediments of volcanic origin include all fragmental materials ejected from volcanoes. They consist of fine volcanic dust, ash, cinders, and bombs (page 314).

Sediments of magmatic origin are not extensive. They represent dissolved substances that were transported from within the earth by the heated waters associated with magmas. Much of the material reaches the surface in hot springs where it may be deposited on land, as in Yellowstone Park, or the springs may discharge on the floor of the sea and add their dissolved load to the sea water.

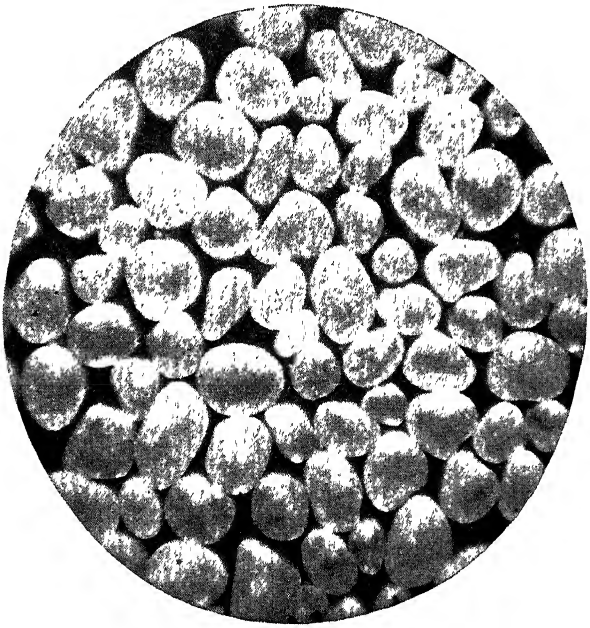

FIG. 275.—Photomicrograph of sand grains from the St. Peter sandstone showing the degree of rounding and the frosted surfaces of quartz grains. (Twenty times natural size.)

Nature of Sedimentary Materials.—The materials that accumulate to form sediments are solid fragments or materials that were dissolved and carried in solution to the basins of deposition. The chief agents of transportation are water, ice, atmosphere, gravity, and organisms. The solid fragments are commonly classified on the basis of their size, as boulders, cobbles, pebbles, granules, sands, silts, and clays (Figs. 274, 275). Several size classifications have been proposed; that given on page 254 is commonly accepted.

Classification of Fragments

Name of Fragment	Diameter in Millimeters
Boulder	256 or more
Cobble	64 to 256
Pebble	4 to 64
Granule	2 to 4
Sand	1/16 to 2
Silt	1/256 to 1/16
Clay	Smaller than 1/256

STRUCTURAL FEATURES OF SEDIMENTS

Stratification.—The arrangement of sediments in layers or strata is their most distinctive structural feature. The stratification may be due to some change in color, to different sizes and kinds of material, or to some interruption in deposition, permitting changes to take place before more material was deposited. The difference in character of sediments may result from (1) variations in currents, (2) seasonal changes, (3) climatic changes, (4) fluctuations of sea level, or (5) marked changes in the types or number of organisms. The individual layers range in thickness from a fraction of an inch in some clayey muds to many feet in coarser sediments. If the bedding planes are near together, the sediment is "thin-bedded"; and if they are far apart, it is "massive" or "heavy-bedded." When the beds are laid down, they are generally nearly parallel to the surface over which they are deposited. As a rule they are approximately horizontal. At many places, however, the surfaces of deposition are undulating, and inclined stratification results. Sediments may be deposited in orderly sequence upon surfaces inclined as much as 30°. The steepest slopes formed by deposition are found in small bodies of water or protected bays where there is slight agitation and very limited spreading of sediments.

Cross-bedding.—Sediments that show parallel bedding at an angle to the planes of general stratification are *cross-bedded* (Fig. 276). Coarse sediments, such as pebbles and sand, are more likely to show cross-bedding, although some sandy shales and limestones are cross-bedded. Wherever steep slopes are produced by the rapid deposition of sediments, whether in rivers, lakes, or the sea, as at the front of a delta or offshore bars, barriers, etc., inclined stratification occurs. Succeeding beds again may assume a horizontal position, and the cross-bedded layers thus may be interstratified with horizontal ones. Wind-laid deposits, such as sand dunes, characteristically are cross-bedded, and in them also cross-bedding is produced by rapid deposition at the front of steep slopes of the growing deposit.

Lenticular Beds.—Massive beds, such as sandstones or limestones, at many places decrease rapidly in thickness and may be seen even to "pinch out" when traced along the face of a cliff (Fig. 277). Lenticular bedding is common especially in the deposits made at the outlets of rivers or in stream deposits generally, but it is not limited to such localities. At the margin of one kind of sediment, where it grades into another kind, the change often takes place through an interbedded area in which the ends of the beds are lenticular or pinched out. Thus the bedding evidently is not parallel, although it is not to be classed as cross-bedding. Irregularities in bedding are to be expected and may be brought out by comparing outcrops that are separated by several miles, although such features may be found also in adjacent outcrops.

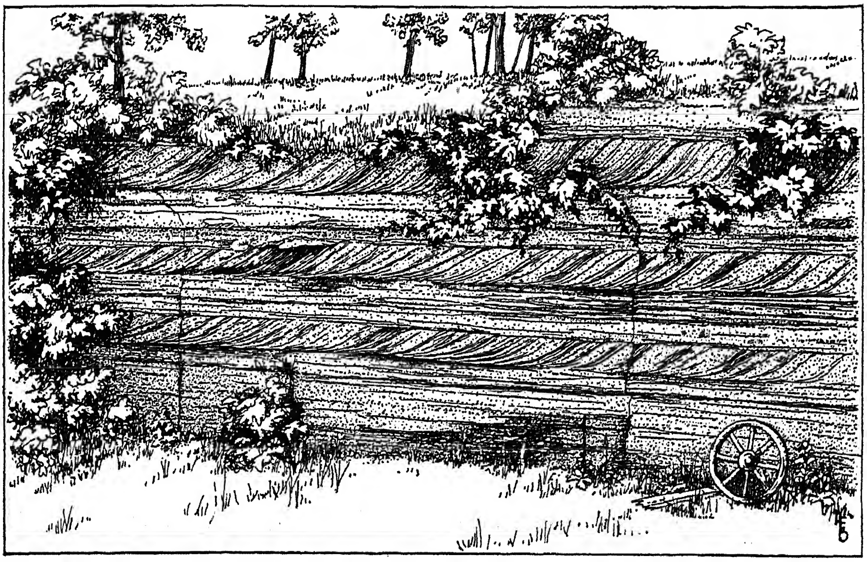

Fig. 276.—Cross-bedding in sandstone strata, near Hawesville, Kentucky.

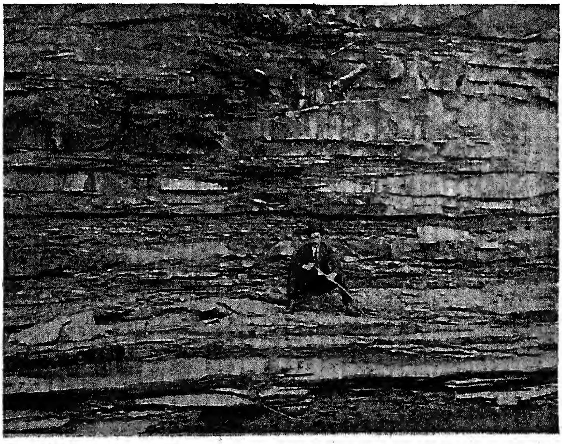

Fig. 277.—Lenticular beds of sandstone in the Berea grit at Gahanna, Ohio.

Mud Cracks.—Mud cracks are due to the shrinkage of the mud on drying (Figs. 278, 279). This results in the removal of some of the water in which the mud accumulated as a sediment. Where formed in marine beds, mud cracks are a marked characteristic of sediments accumulating at or near the shore line or where the muds are exposed to the drying effects of the sun part of the time. Some mud cracks are several inches wide and ten times as deep. Such cracks, when the mud is thoroughly dried out, may remain open for several years even in regions

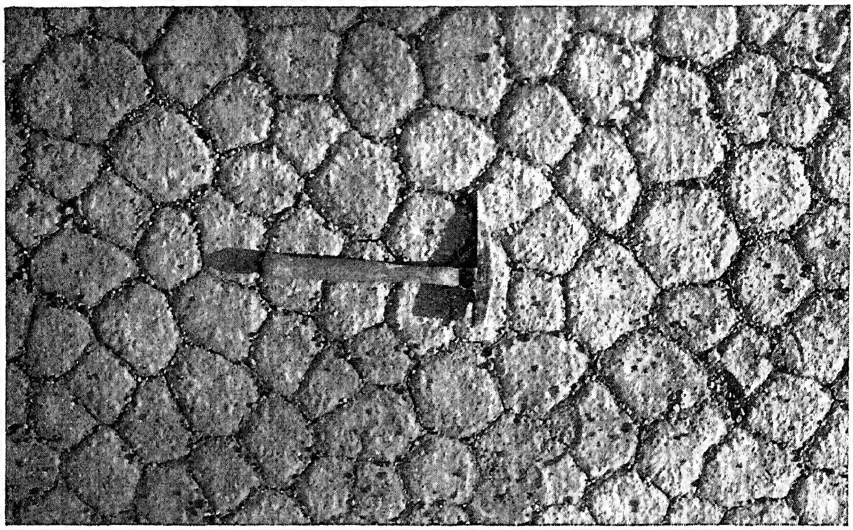

FIG. 278.—Mud cracks in Playa Lake sediments in the Great Basin region of Southwestern U. S. (*Photograph by Chas. Erdman, U. S. Geol. Survey.*)

FIG. 279.—Mud cracks on the flats adjacent to Lake Calhoun, Minneapolis, Minnesota.

of moderate rainfall. Where water again accumulates over them and sedimentation is renewed, the cracks are filled by this later and perhaps different kind of sediment. Such sediments, when thoroughly indurated, preserve the mud cracks as a feature of the solid rock. Where a mud-cracked area is drying rapidly, under certain conditions the patches thus blocked out on the upper surface may peel and roll up into cylinders. These, if preserved in the oncoming sediment, may resemble pebbles and produce a dessication conglomerate.

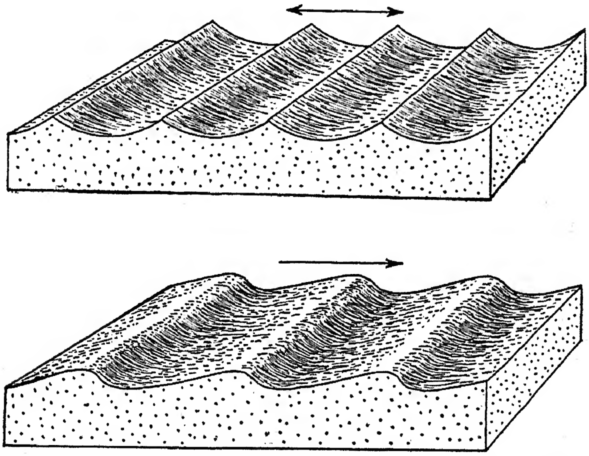

Fig. 280.—Diagrams showing the difference between oscillation (above) and current ripple marks (below). The arrows indicate the direction of movement of the water.

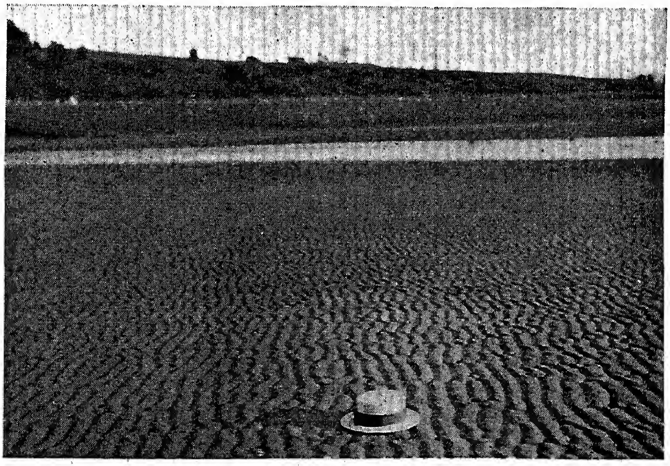

Fig. 281.—Ripple marks left by receding tide at Windsor, Nova Scotia. (*Courtesy E. M. Kindle, Can. Geol. Survey.*)

Ripple Marks.—Ripple marks usually are formed by the drag of the waves as they strike and travel over the bottom in relatively shallow water (Figs. 280, 281, 282). They consist of a series of small, almost equally spaced ridges of sand or other fine sediment with rather sharp crests. The slopes of the sides show the type of motion which formed them. Oscillatory or undulatory movements give rise to ripples with symmetrical sides, whereas currents give rise to ripples with asymmetrical sides, the long slope being in the direction from which the current came. Under similar conditions, as the undulation dies out in deeper water, the crests of the ripple marks are more closely spaced, although the coarseness or fineness of the sediment composing them also may be a factor in the spacing. Waves in shallow water may be too violent at the contact of water and sediment to produce ripples. Their action is destructive, and any ripple marks that may have been formed during a more quiet sea are obliterated. Sand waves or giant ripple marks have been observed. These usually are formed in narrow bays by tidal currents. The wind forms ripple marks in land deposits such as sand dunes, and these resemble some of those formed in the water.

Fig. 282.—Ripple marks preserved in the consolidated transitional beds at the base of the Beekmantown dolomite, Perth, Ontario. (*Courtesy M. E. Wilson, Can. Geol. Survey.*)

Rill Marks.—As the tide retreats or the storm waves die down, the water left in the sand of a beach finds its way back to the sea and produces little rivulets which branch again and again like the distributaries on a delta plain. These are known as rill marks. Commonly they are observed on the beach sands and frequently are found in the consolidated sediments where at places they have been mistaken for plant impressions.

Wave Marks.—Wave marks are formed on the sloping sands of a beach by the outer margin of the spent wave. They may be indicated by smooth-lying fine sand grains at places partly covered by fragments of shells and bits of mica. Seaward from the marginal mark often may be found various patterns produced by the returning water. Of these some have an imbricated design like overlapping scales.

Raindrop Impressions.—Raindrop impressions are produced by rain or by the splash of wave action or by water dripping from vegetation, and they may be preserved in the mud along the shore or on other mud flats. For their preservation probably it is necessary for partial drying or hardening of the mud to take place before the mud is covered and then that it be covered quickly by another layer. Such marks are not commonly found, but at places excellent examples are preserved.

Places of Deposition.—Eventually most sediments find their way to the sea, although some find temporary lodging in various continental environments such as flood plains of rivers or in lakes. Most realms of accumulation pass laterally into each other either as the gradual transition between the waters of the deep and shallow sea or as the flood-plain environment which passes into that of the delta, and the latter in turn into the shallow-water marine realm. Likewise, both flood plains and deltas have shallow lake basins in which sediments accumulate.

The basins and other places where sediments are deposited are continental and marine. Since these two realms come into contact along the sea coast, it follows that an area adjacent to the shore line has some of the characteristics of both. Thus a third division of mixed continental and marine conditions may be included. The various places where sediments accumulate may be tabulated as follows:

Continental	Mixed continental and marine	Marine
Terrestrial:	Littoral	Shallow sea
Desert	Lagoon	Intermediate sea
Glacial	Estuary	Deep sea
Fluvial:	Delta	
Piedmont		
Valley flat		
Lake		
Swamp (Paludal)		
Cave		

CONTINENTAL SEDIMENTS

Desert Deposits.—At the present time approximately 11,000,000 square miles of arid desert regions exist on the surface of the earth. The sediments of these regions accumulate by wash from upland slopes, by intermittently torrential streams, by deposition from waters of playa and saline lakes, and by the deposition of wind-blown sediments. Most desert sediments are more or less etched, frosted, and polished. This is true especially of the coarser lag materials over the rocky desert platforms. The valley and gully deposits are composed of coarser detritus that extends up the valleys into the highlands and down the valleys to alluvial cones and fans. These coalesce laterally to form piedmont slopes so that many desert mountains appear to rise out of gravel deposits. The fine sands and silts of the playa and salt lakes dovetail with the dune and piedmont deposits. Stratification is conspicuous in the laminated clays of the lakes and playas, whereas most of the gravels on the piedmont slopes are unstratified. The eolian sediments are characterized by wedge-shaped, cross-bedded units.

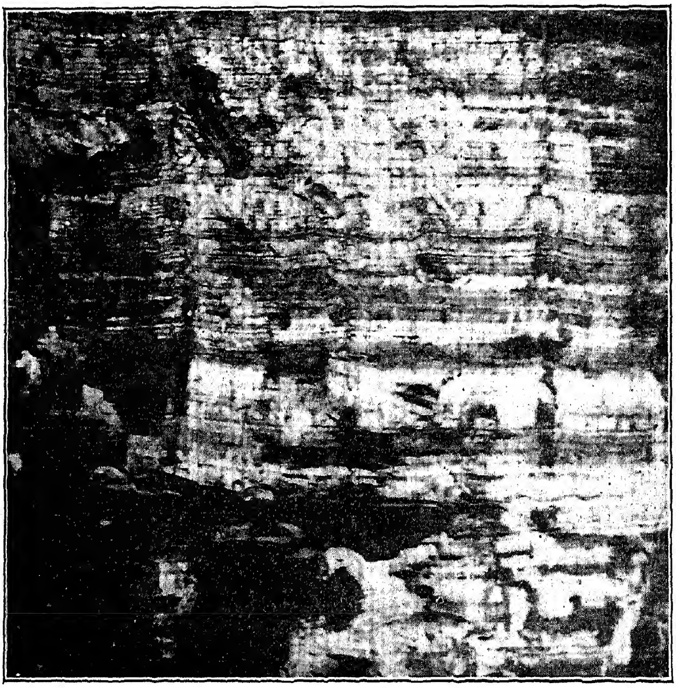

FIG. 283.—Laminated clay, Wrenshall, Minnesota. Indurated clays form shale. (*Photograph by Grout.*)

Glacial Deposits.—Glacial sediments are described in Chapter X. Ice-deposited sediments are unstratified, unsorted mixtures of coarse and fine sediments that are bordered laterally by and are interbedded with stratified sediments of eskers, outwash plains, and glacial lakes. The fine-grained sediments of the glacial lakes may be varved (Fig. 283). An individual varve may consist of clay and silt in various proportions, or it may be all silt or nearly all clay. The coarser the material the thicker will be the summer as compared to the winter layer. Glacial sediments commonly rest with a sharp contact on striated and grooved rock surfaces.

Fluvial Sediments.—Fluvial sediments are those deposited by flowing water (Fig. 284). They represent the products of aggradation on piedmonts, valley flats, and the upper surfaces of deltas. Piedmont sediments accumulate about the bases of mountains as a result of soil creep, rain wash, rock streams, mud flows, and intermittent streams. The present extent of piedmont deposits in western United States has been estimated as being equal in area to that of the mountains above them. In the Cucumonga district of California the deposits are more than a thousand feet thick. They are composed of boulders, cobbles, gravel, sand, and silts. The sediments are poorly sorted and are indistinctly bedded, with the coarser fragments nearer the base of the mountains.

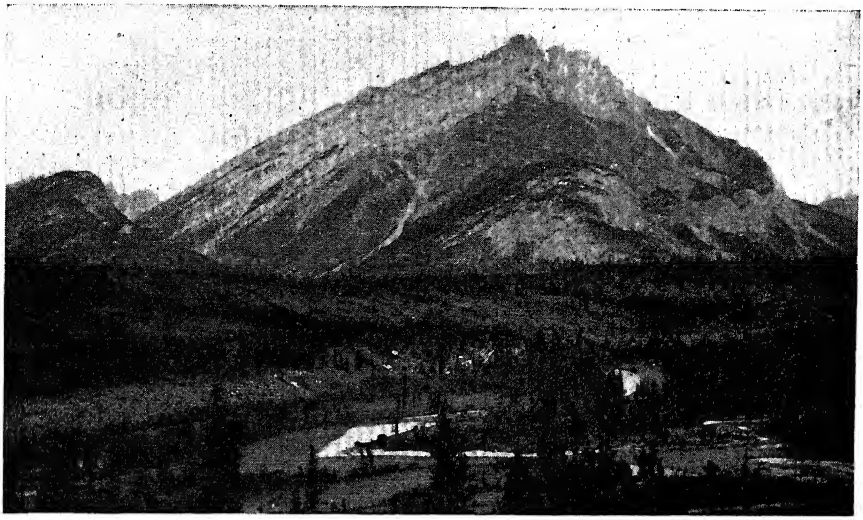

Fig. 284.—Alluvial terrace sediments (foreground) in the valley of the Bow River at Banff, British Columbia.

The sediments of the valley flats differ from those of the piedmonts in that they show better sorting and stratification, fewer large fragments, and more organic matter. Since most streams alternately aggrade and degrade, the sediments have but temporary lodgment. Many flood plains contain lakes and swamps, both of which shift their positions as the streams change the location of their channels over the valley flats. There thus results a dovetailing of lake clays and silts with swamp mucks and peats, and these in turn alternate with sand and gravel transported by the streams. Many of the sedimentary rocks of the Great Plains region were originally deposited as fluvial sediments.

Lake Sediments.—The processes operating in the filling of lake basins have already been discussed (page 218). Lake sediments consist of marl, tufa, peat, iron and manganese oxides, iron carbonates, gravels, sands, clays, salt, gypsum, and other saline products of evaporation.

Swamp Sediments.—It has been estimated that the area of swamps in the United States is 100,000 square miles. They are extensively developed along the coasts of lakes and of oceans, especially in the tropical and subtropical regions. In higher latitudes swampy basins are most numerous in the regions that were recently covered by continental ice sheets. Since vegetation usually covers the entire surface of swamps, the drainage is checked and evaporation is retarded. Where the drainage is checked appreciably, the level of the water rises, and the swamp expands laterally. The swamps connected with streams usually have considerable inorganic matter incorporated with the organic. As the organic residues decay, black mucky deposits are formed. In swamps that do not receive waters carrying mud and sand, clean peat is deposited. In the latitude of the United States the chief peat-forming plants are mosses, ferns, canes, sedges, grasses, rushes, and small evergreen trees.

Cave Deposits.—Cave deposits are, for the most part, the products of precipitation from underground water. The characteristics of the deposits are discussed on page 87.

MIXED CONTINENTAL AND MARINE SEDIMENTS

The sediments that accumulate where the oceans and the continents meet are a mixture of material derived from the land and from the sea. They accumulate along the shore zone and in the lagoons and estuaries. Conditions of deposition in the littoral zone are not everywhere the same. Some shore zones consist of bare rocky platforms; others are nearly vertical sea cliffs; and still others are composed of gravels, sands, muds, shells, and shell fragments. These sediments grade into each other alongshore and grade seaward by imperceptible stages into the offshore marine deposits. The sediments of the shore zone are derived mainly from the shore by wave action. The waves are aided by frost, by undercutting, and by the wind. The work of the wind is more important, however, in generating waves and currents that carry sediments to the beaches. The materials of the beach vary with the source of supply and the vigor of the wave action. Upon a boisterous or surf-beaten coast the materials may be boulders or large cobblestones. Where the supply of finer materials is extensive, even on exposed coasts, the material may be pebbles, or it may be sand. On rocky coasts beaches of boulders and cobblestones commonly form at the heads of indentations, although at places sand may occupy such positions. These *pocket beaches* are found along the coast of California at Carmel, at La Jolla, and at many other places. They are merely lodgment places in which the rock fragments are ground to fine particles and from which they are finally swept out to sea by the undertow. The grinding process is caused by the surf rolling up and down the beach dragging the boulders, cobblestones, and pebbles back and forth over each other and over the rock-shod bottom. As agitation ceases and the sediments finally are deposited, they are graded in the order of size from the shore outward.

In the marginal lagoons the waters range from fresh water to waters with a salinity greater than that of the adjacent sea. The sediments that accumulate there likewise exhibit a considerable range. Land-derived sediments are brought by streams and wind; marine sediments are brought by currents from the sea; and organic and chemical precipitates are produced from the salts in solution. Calcareous marls are precipitated by plants and invertebrate animals and to some extent by direct chemical precipitation. In stagnant lagoons the activity of bacteria leads to the formation of hydrogen sulfide, which causes the precipitation of black iron sulfide in the accumulating sediments. In such black muds carbonate shells dissolve and are replaced by iron sulfide. At places where there is extensive evaporation the salinity may become so great that beds of salt and gypsum are deposited.

Delta sediments also accumulate in the mixed continental and marine realm (page 151).

MARINE SEDIMENTS

The realms of marine sedimentation include the shallow epicontinental seas, the continental or intermediate slopes, and the deep sea (Figs. 243, 244).

Shallow Deposits.—The shallow sea is that portion of the ocean basin extending from the low-tide level to the depth of 100 fathoms. It includes the major portion of the continental shelf, together with such epicontinental seas as the Baltic Sea and Hudson Bay. The distribution, extent, and depths of the shallow sea varies with the nature and extent of diastrophic movements. Where the sea level rises, the deeper portions of the shallow sea are added to the intermediate slopes, and the lowlands along the coast are flooded so that the former littoral zone is added to the area of the shallow sea (Fig. 289). During the geologic past when large areas were reduced to low peneplanes, a rise of sea level extended the shallow sea far into the interior of the continents and added many thousands of square miles to the realms in which shallow-water deposits accumulated. Most of the marine sedimentary rocks of the continents of today accumulated in such basins.

The shoreward portion of the shallow sea bottom lies within the range of wave and current action. Where such currents are generated the sediments are sorted so that the coarser materials are deposited near shore and grade into finer deposits seaward. There are many exceptions, however, to this generalization, for along low shores with few streams, fine muds and calcareous sediments may accumulate far up on the beach. Recent studies of sediments from the continental shelf along the eastern coast of the United States show that the distribution of sediments is extremely irregular (Fig. 285).

Where the supply of land detritus is small, the sediments on the sea bottom may consist chiefly of the remains of organisms and of chemical precipitates (Figs. 271, 272). Some precipitates may be formed in the sea by reactions between constituents of different origin which produce insoluble compounds. It is not known to what extent calcium carbonate is precipitated by a purely inorganic process. It is doubtful whether simple precipitation takes place to any appreciable extent in the ocean under usual conditions, and probably most marine calcareous sediments are of organic origin. Such sediments form in comparatively shallow water, where lime-secreting organisms live in abundance but where there is relatively little sediment being carried in from the adjacent land. Partial disintegration of the shells, due to the grinding by waves or to

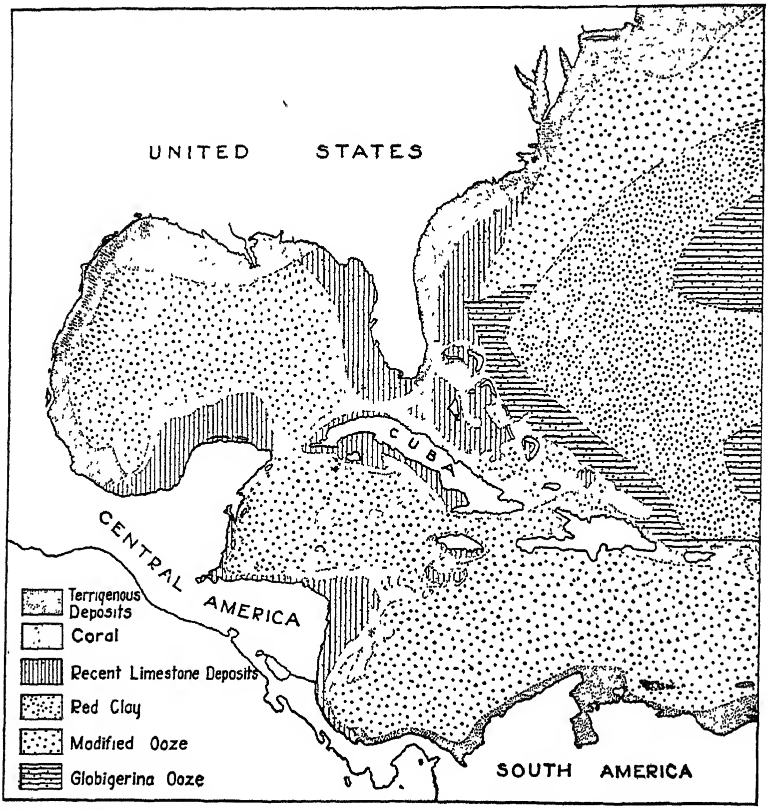

FIG. 285.—Map showing the distribution of bottom deposits along the western margin of the Atlantic Ocean basin in the vicinity of the Caribbean Sea, the Gulf of Mexico, and the southern coastal plain of the United States. (*Based on a map by Agassiz.*)

solution, may destroy their original structure before they have been transformed into solid rock, or recrystallization may destroy remnants of forms that are left. The resulting unfossiliferous rock might be mistaken for a direct chemical precipitate.

The calcium carbonate content of the offshore sediments is influenced by temperature, depth, salinity, degree of saturation of the water with calcium carbonate, and the activity of living organisms. It increases as the surface salinity rises. The rate of rise in carbonate is greatest between salinities of 34 and 36 parts per thousand.

The structural features of offshore marine sediments are variable. Deposits near shore are usually lenticular beds, with much cross-bedding and a great range in size of particles. Ripple and current marks have equally great variation in trend and in extent of development. Where the sea floor has steep slopes the sediments may slump and develop crumpled and irregular bedding planes. In deeper waters the strata are more nearly uniform, and the stratification is distinct. Some chemical sediments show well-defined seasonal laminations.

Intermediate and Deep-sea Deposits.—Beyond the 100-fathom line the ocean bottom descends rather abruptly to the sea floor with an average depth of 2½ miles below the surface. This descent is the continental slope or outer margin of the true continental mass. It begins at an average distance of about 10 miles from the coast; hence, in general, it is covered by those fine sediments of land origin which remain in suspension for a long period. These collectively are called the blue muds and owe their color to the presence of organic matter and to the deoxidized condition of the iron. Landward they grade into the shallower water deposits, and seaward they pass into the oozes and red clay of the abysmal depths. The blue muds probably cover 15,000,000 square miles of the ocean basin. They have been encountered at distances from land as great as 200 miles, and out from the mouths of great rivers, such as the Amazon, they extend a distance of 1,000 miles. Volcanic mud, pieces of pumice, and scoria have been picked up by dredges more than 200 miles distant from the volcanic islands of Hawaii. Land-derived vegetation is strewn over the sea bottom at places even at a distance of several hundred miles from the land and in 200 fathoms of water. Such vegetation was found off the coast of Central America on both the Atlantic and the Pacific sides, where nearly every haul of the *Challenger* dredge brought up fruits, seeds, leaves, twigs, branches, and parts of the trunks of trees. This organic matter contributes to the deoxidizing agents of the blue-mud zone, but probably little is preserved in fossil form.

With increasing distance from shore the land-derived materials assume less and less importance. In the deep abyss many sediments are of volcanic, pelagic, glacial, and meteoric origin. Because of the nature of the environment, the sediments are altered before they reach the floor of the deep sea. At great depths the pressure is tremendous. It rises at the rate of more than a ton per square inch per mile of depth, so that at a depth of 4 to 6 miles the weight on each square inch is 5 to 7 tons. The temperature is approximately 35°F. at all times; and since currents and waves of the near-shore type do not exist, there is no appreciable motion of the water. There is no light other than that emitted by the phosphorescence of some deep-sea organisms. Fewer organisms exist than in shallower waters. The chief organic sediments consist of the hard parts of organisms which lived in the upper lighted waters. These surface-dwelling forms are chiefly simple types of plants and animals collectively called the *plankton.* They consist of small Mollusca, Foraminifera, and Algae, which secrete calcium carbonate, together with diatoms and radiolarians, which secrete siliceous skeletons. When these organisms die, their remains sink to the bottom where the undissolved residue, together with volcanic, meteoric, and other dusts form ooze or slimy deposits that accumulate very slowly. These oozes are named according to the most abundant remains composing them. Thus there are the globigerina oozes, the pteropod oozes, the diatom oozes, the radiolarian oozes, etc., but they all grade into each other (Figs. 271, 272).

Globigerina ooze is by far the most extensive, and it is now forming over areas of many millions of square miles. The radiolarian ooze is found in some of the deepest parts of the sea.

Owing in part to the increase, with depth, of the carbon dioxide in sea water, the percentage of calcium carbonate in the bottom deposits decreases in general with depth. This is because the carbon dioxide and water form a weak acid which dissolves the calcium carbonate. Considering the oozes and muds of various origin together, the calcium carbonate content of the deposits on the sea bottom is as follows:

VARIATIONS WITH DEPTH IN CALCIUM CARBONATE OF BOTTOM DEPOSITS
(After F. W. Clarke)

Fathoms	Per Cent	Fathoms	Per Cent
Under 500	86.04	2,000 to 2,500	46.73
500 to 1,000	66.86	2,500 to 3,000	17.36
1,000 to 1,500	70.87	3,000 to 3,500	0.88
1,500 to 2,000	69.55	3,500 to 4,000	None

In the greater depths of the ocean the bottom is covered by a very fine red clay which is composed of the insoluble portions of the plankton shells or volcanic ash and the meteoric dust from the heavens. Young[1] states that approximately 20,000,000 meteorites enter our atmosphere

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[1] YOUNG, CHARLES A., Manual of Astronomy, p. 464, 1902.

daily. Most of them are disintegrated in their passage through the air and settle as the finest dust over land and sea. If the average weight of these is ¼ ounce, the total weight of such material reaching the earth yearly would be about 57,000 tons. When three-fourths of this is scattered over the whole ocean basin, the amount is understood to be trivial. It is evident, therefore, that sedimentation on the deep-sea floor is exceedingly slow. The rate at which it is taking place may be appreciated from the fact that dredging on the sea floor often brings up the inner ear bones of the whale; these are the most resistant parts of the whole skeleton and have thus accumulated on the sea bottom through many generations of whales, and yet they are so thinly covered by the red clay that frequently the dredge will bring them up, sometimes as many as 90 in a single haul. At Station 285, in the South Pacific, a single haul of the *Challenger* dredge brought to the surface 1,500 sharks' teeth, many of them representing extinct species, in addition to immense numbers of very small teeth and fragments. Evidently they had been accumulating for a very long time, and although manganese coated, they were practically without sedimentary covering. The red clay occupies about 50,000,000 square miles of the sea bottom, and most of the known area is in the Pacific Ocean.

Although the chalk beds of England and France are chiefly the remains of Foraminifera and a large part of the Monterey series at Lompoc, California, consists of the remains of diatoms, it is doubtful whether these or any beds found on the present continents correspond to the deep-sea oozes or the red clays. Steinmann reports them in the Alps and the Apennines, and others consider some of the Radiolarian cherts of the Northern Appalachian region to be deep-sea deposits; but even if these should prove to be true abysmal deposits, the total known land area would be small.

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CHAPTER X

SEDIMENTARY ROCKS

Origin and Characteristics.—Sedimentary rocks are derived from sediments which have undergone compacting and cementation or binding together of their particles. They are bedded or stratified rocks because the materials which compose them are laid down in layers or strata. The stratification in most of these rocks is evident from the planes of separation of beds. Usually such planes are due to thin films of sediments of a texture finer than that composing beds adjacent, but differences in composition of successive beds may produce the same effect. Sedimentary rocks record the conditions of sedimentation at the time they were deposited. Sandstones show ripple marks, wave marks, cross-bedding, tracks, burrows, etc., just as do sands of the seashore today. The same is true of other kinds of sedimentary rocks, and the interpretation of these ancient sediments is made through a knowledge of what is taking place under similar conditions today.

Consolidation of Sediments.—As sediments increase in thickness, the lower beds are pressed together by the weight of the overlying beds and are rendered more compact. Lateral pressure or any other movement may drive the rock particles closer together, decreasing the pore space and consolidating the rock. The most important consolidating process, however, is cementation. The percolating ground waters carry calcium carbonate or other cementing materials into the pore spaces of the fragments and there deposit them, cementing the grains or larger fragments into a solid mass. The common sedimentary rocks are the results of this process. No distinct line is drawn, however, between the loosely aggregated sediments and the compact hard rock that results from the consolidating process. In general, the older sediments are more highly consolidated than the younger ones, because there has been a longer time for cementation. This is not always true, however, for some of the ancient sediments are little more than beds of gravel, sand, or mud; either they were never consolidated, or the cementing material has been dissolved and removed.

Types of Sedimentary Rocks.—The crust, or outer 10 miles, of the earth is estimated to have the following composition: 95 per cent igneous and 5 per cent sedimentary rocks. Of the sedimentary rocks about 83 per cent are shales, about 10 per cent sandstone and conglomerates, and 7 per cent limestones. Sedimentary rocks may be divided into the *clastic* and *non-clastic* types. Clastic rocks are those composed of the fragmental materials resulting from the breaking up of other rocks. The most common ones originate through the weathering, transporting, and depositing processes. These give rise to the conglomerates, the sandstones and the shales. The non-clastic sedimentary rocks in general are derived through the agency of some form of life (biochemical) or

A TABLE OF COMMON SEDIMENTARY ROCKS

	Rock	Material	Composition and remarks
Clastic Rocks	Arkose	Derived from disintegrated granitoid, acidic igneous rock	Mainly feldspar and quartz
	Breccia	Angular fragments and cement	Cemented angular fragments of any kind of rock
	Conglomerate	Rounded pebbles and cement	Cemented pebbles
	Pebbly sandstone	Sand with scattered pebbles	Sand grains and pebbles
	Grit	Coarse sand, usually sharp	Usually cemented coarse sand
	Sandstone	Subangular to rounded grains, usually quartz	Usually SiO_2. Cemented beds of common sand
	Greensand	Sand grains and glauconite	Sandstone conspicuous with glauconite
	Shale	Clay partly indurated	Thin, flaky, or platy
Chemical and Biochemical Precipitates (Non-clastic)	Limestone	Fine particles or shells usually finely crystalline	$CaCO_3$. Effervesces in cold HCl
	Dolomitic limestone	Slightly magnesian with or without shells	When scratched may effervesce mildly in HCl
	Dolomite	Fine particles mainly crystalline. May contain shells	$CaMg(CO_3)_2$. Effervesces in hot HCl
	Hydraulic limestone	Argillaceous limestone sometimes called water lime	Makes a cement that sets under water
	Lithographic limestone	Very fine-grained calcareous material	$CaCO_3$. A stone used for making lithographs
	Travertine	Precipitate from springs	$CaCO_3$. In caves, stalactite and stalagmite
	Chalk	Fine calcareous sediment usually of very small shells	Usually a fine white or gray limestone of light weight
	Shell marl	Shells and a shelly calcareous paste	$CaCO_3$. Usually fresh-water shells
	Coquina	Shells and shell fragments	A limestone made up of conspicuous shells and shell fragments
	Oolite	Small spheroidal granules	Calcareous or siliceous
	Chert	Compact opaline silica	SiO_2. Lenses and nodules in limestone. Some massive beds
	Diatomite, or diatomaceous earth	Fine siliceous framework, or frustules, of diatoms	Light in weight and usually white
	Lignite	Preserved and compressed plant material	Early stage in coal formation
	Coal	Carbonaceous plant material with plants usually inconspicuous	Usually means ordinary bituminous coal
	Gypsum	Chemical precipitate due to evaporation	$CaSO_4 + 2H_2O$. Without water it is anhydrite
	Salt	Chemical precipitate due to evaporation	NaCl. Clear crystals are halite
	Bedded iron ore	Oxides, carbonates, and sulphides. Clastic and non-clastic	Usually hematite or limonite. Some is pyrite or siderite

are chemical precipitates. They are derived only indirectly from other rocks through their decomposition and redeposition. Their materials existed originally in the older rocks, but by weathering and solution they have been separated, and, in general, running water has carried them to the sea to form the salts of the ocean.

Characteristics of Common Sedimentary Rocks

Arkose.—Arkose is a rock composed of the residue of disaggregated, granular, acid igneous rocks. Its constituents are mainly unaltered feldspar and quartz. It is a sandstone in which fragments of orthoclase feldspar are abundant. If the original rock was basic and the feldspar plagioclase, the derivative is called *graywacke.* Arkose deposits are common among the land-laid formations and especially among those accumulating in arid or semiarid mountainous regions, where the chemical processes of decomposition are slow and where other processes of weathering are more effective.

Fig. 286.—Limestone breccia. Fragments of white limestone are cemented by darker calcium carbonate deposited by solutions.

Breccia.—A rock composed of the cemented angular fragments of other rocks is a *breccia.* It is evident that the angular constituents of breccia have not been transported by water far from the source of the material (Fig. 286). Breccias are common along fault zones or in ancient talus accumulations. They grade into conglomerates as the fragments show signs of rounded angles due to transportation by water.

Conglomerate.—Conglomerates are gravels that are held together by some kind of cement. Many conglomerates are made up chiefly of quartz pebbles, because quartz is the most common mineral that possesses great resistance to disintegration and wear. Flint, chert, and jasper also are common, although the pebbles of a conglomerate may be made up of any kind of rock fragments. Thus there are limestone conglomerates and also shale conglomerates. Unassorted pebbles, such as those found in an ordinary gravel bed of the glacial drift, especially where there is a high percentage of calcium carbonate in the associated drift, may be cemented into great masses of rough conglomerate which look like weathered concrete.

Sandstone.—A sandstone is a bed of sand cemented to form a coherent mass. The cement may be the light-colored calcium carbonate or the red, yellow, or brown iron oxide. If quartz sand is thoroughly cemented by silica, the rock is quartzite. Sand is made of grains smaller than pebbles. The grains may be subangular or well rounded. Their surfaces may be pitted owing perhaps to impact in transit, or they may be frosted like the surface of ground glass. Thus they tell something of the history through which they have passed as fragments. Most sands are made up chiefly of quartz fragments, but sands made of fine grains of olivine are found along the Bay of Naples, sands of calcium carbonate are found at Bermuda, and sands composed of gypsum are found in New Mexico. Certain sands are composed of magnetite, and others partly of tin oxide or of gold, but all such sands are unusual. Sandstones grade through coarse-grained sands into conglomerates and through fine-grained sands into shales. Sandstones that contain an appreciable amount of calcium carbonate are calcareous sandstones; if they contain clay, they are argillaceous sandstones. Occasional pebbles may occur in sandstones; these are strung along a bedding plane or occur here and there in the sand. Such a rock is not a conglomerate but a pebbly sandstone. A coarse sandstone, especially if the grains are sharp, is a *grit.*

Greensand.—Grains of glauconite may be common to abundant in a sandstone. In some sandstones they are more abundant than quartz grains. Such rock is called greensand, or a glauconitic sandstone. It is common among some of the older sediments and is now forming in the ocean.

Shale.—Shale is composed of compacted or cemented beds of mud or clay. It includes the finest products of mechanical rock decay, and it is swept farther out to sea than any other clastic sediment. Shales usually are thin bedded, showing frequent changes in the fineness of materials composing them. Such changes in grain may be due to seasonal changes, or they may represent differences in rainfall or some other change that affected the amount or character of sediment brought down by streams. Shales that contain sand are arenaceous, those that contain calcium carbonate are calcareous, and those containing iron are ferruginous. Those containing large amounts of organic matter are bituminous shales. The latter usually are black, and some grade into beds of coal. Some of the ancient shales remain as beds of clay and differ from the deposits originally formed only in that they have been pressed together slightly.

The weathering of limestones may give rise to beds of clay which represent the insoluble residue of the limestone. Red clay beds may be so formed. When covered by later sediments, these residual clays mark disconformities or ancient erosion surfaces.

Thin-bedded limestones or sandstones often are referred to as shaly, and many of them grade into shale.

Fig. 287.—A slab of fossiliferous Hamilton limestone from western New York. (*Photograph by Hardin, U. S. Geol. Survey.*)

Limestone.—Limestone, the most abundant non-clastic sediment, consists of a solid rock made up of the shells and the skeletal materials of lime-secreting plants and animals (Fig. 287). Such organisms extract this material from sea water to form their hard parts. The spaces between the shells are filled by fine calcareous materials resulting from the grinding action of waves. Limestones are formed in relatively shallow water where life is abundant and where the neighboring land areas are too low to contribute large quantities of clastic sediments. Living bacteria may cause the precipitation of calcium carbonate in sea water, and under exceptional conditions limestone may be deposited by chemical precipitation, but the greater number of limestones, whether fossiliferous or not, probably are composed of material that once formed organic remains.

Limestones may grade into shales by the addition of clay to the calcareous sediments. They may grade into calcareous sandstones in like manner by the addition of sand and through them into ordinary sandstones.

Dolomite.—If a large part of the calcium in a limestone is replaced by magnesium, the rock is dolomite. Dolomitization is a common process in limestones of all ages and often is accomplished during the process of sedimentation or by substitution of magnesium for calcium before the limestone is covered by later beds.

A limestone in which no such change has taken place is rare. If there has been only a small amount of replacement (10 per cent to 20 per cent), the rock usually is referred to as a dolomitic limestone. Dolomites and dolomitic limestones are less soluble than ordinary limestones, hence endure longer on exposure. On weathering of limestone, calcium carbonate is dissolved and removed more rapidly than magnesium carbonate, and the proportion of magnesium carbonate is increased.

Chalk.—Chalk is a special type of limestone usually composed of small shells, or of their fragments, cemented together. Foraminifera shells, or tests, constitute a large part of the material, but shells of other organisms also are commonly present.

Chalk usually is soft, porous, white or gray, and some of it is massive in appearance. The chalk cliffs of Dover, England, are an example. Some of the chalks of the southwest, particularly of Texas, grade into resistant beds that are as well indurated as ordinary limestones, and such beds are found capping buttes and mesas of the region.

Marl.—The porous masses of shells and shell fragments that accumulate on the bottoms of many fresh-water lakes are shell marls. Large amounts of marls are formed by the lime-secreting alga, Chara. The best known example of Chara marl occurs at Pyramid Lake, Nevada. The term marl is used also to designate certain marine sediments presumably formed at the outer margin of the shale mud, in which clay and finely divided shell fragments are present. The term is applied also to soil in which clay and calcium carbonate are present in about equal amounts. Marine muds that are composed chiefly of calcium carbonate are called marls by some, but this practice is not general. The greensands of New Jersey usually are called greensand marls.

Coquina.—Coquina is a limestone composed of loosely aggregated shells and shell fragments. This term usually is applied to the more recent deposits of cemented shell heaps, such as those forming off the coast of parts of Florida, but it is applied also to similar shell masses belonging to much older formations in which the mass is well consolidated.

Oölite.—An oölite is a rock in which the particles consist of small concretion-like particles resembling fish roe. The term is generally applied to the texture of the rock, for there may be siliceous, calcareous, and ferruginous oölites. Commonly a sand grain forms the center of the concretion.

Chert.—Chert is a compact, dense, siliceous material that occurs in many calcareous rocks. Silica, as the hard parts of sponges or other organisms, probably was deposited along with calcareous shells when the beds containing the chert were laid down. Later the silica was dissolved and redeposited. In general, the chert is distributed through the beds of rock as rounded nodules, but in certain strata there are extensive chert layers. Flint, a variety of chert, is essentially silica with some water and generally is a somewhat purer chert, although the two names are used interchangeably. In the stone age, man used flint to fashion arrow points, and later he used it with steel to kindle fire. Agate is usually a banded ornamental variety of flint.

Diatomaceous Earth.—Diatoms are minute plants that have siliceous skeletons. They live in great numbers in the sea and in fresh-water lakes. When they die, their siliceous capsules, or skeletons, accumulate to form diatomaceous earth. On many coasts their accumulations form papery gray to white layers interbedded with shales. Such beds have high porosity, and they may contain much water, but when dried they will float on water. At Lompoc, California, and at other places thousands of feet of diatomaceous shales are found.

Coal.—Coal is formed by the compacting and partial decomposition of vegetation accumulated in ancient peat bogs. It is preserved through submergence and by a covering of later beds. The coal bed usually is found in the same location in which the plants, now coal, grew. This is indicated by the flat or unrolled condition of the leaves, the old stumps with roots still penetrating the soil below, and the lack of other sediments mixed with the plant remains. A few coals, however, have been formed from vegetation drifted into bays or estuaries. The alteration of vegetation into peat, lignite, and various other grades of coal is a process usually requiring a long time. The grade of coal produced is dependent upon the kind of material and the amount of alteration, through pressure and heat, that has taken place since the bog deposit accumulated. Coal is included among the sedimentary rocks, but it is merely an accumulation of fossil plants.

Petroleum.—Petroleum is natural oil that exudes from the earth or is pumped up from wells. It is formed by the decomposition of remains of plants and animals that were buried in the muds and limestones that were deposited in the sea. Some of this decomposition probably took place before burial, and the oil was deposited along with the clay and organic remains, but much of it was formed after the organic matter was buried. Petroleum usually is stored in porous sandstones and limestones that are associated with shales. Inflammable natural gas also is formed by decomposition of organic matter, and with salt water it is often associated with oil. Asphalt, grahamite, and other solid bitumens are formed where oil seeps out and hardens in fissures or at the surface of the earth.

Phosphate Rock.—Most igneous rocks contain small amounts of the mineral apatite, a calcium phosphate. Ground water dissolves the phosphate, and plants and animals utilize it in their life processes. Small amounts of phosphate are carried to the sea and are deposited there in beds. Such beds raised above the sea have become available to man, and the phosphate rock is used in large amounts for fertilizer. The largest deposits of rock phosphate in the United States are found in Montana, Idaho, and Wyoming, where they are associated with limestone and cover extensive areas. Valuable deposits are found also in Florida, South Carolina, Tennessee, and Kentucky.

Some phosphate deposits have formed by replacement of limestone, particularly where ground water has leached the waste of animals and dissolved the phosphate, carrying it downward, where it is precipitated by limestone.

Salt and Gypsum.—The composition of sea water is stated on page 226. When sea water is evaporated to dryness, the salts fall out of solution and are deposited. The least soluble salts are deposited first. Calcium carbonate and iron oxide, if present in the water, are the first to be precipitated. Gypsum[1] follows, and often with it some anhydrite is formed. After gypsum, sodium chloride, or common salt, is deposited. The bitter salts consisting of sulphates and chlorides of potassium and magnesium are precipitated last. They are so soluble that they are not always deposited where salt and gypsum form, and if they are deposited commonly they are dissolved again. Great beds of salt and gypsum are interbedded with sedimentary rocks in Texas, Kansas, Michigan, Ontario, Ohio, New York, and many other places. Wherever soluble salts are formed they are likely to be dissolved again unless they are protected to some extent against solution by water. In general, where they are found preserved, muds and clays have been deposited above them.

Red beds, mainly red sandstones and shales, at many places are associated with salt and gypsum. These are red because they contain small amounts of disseminated hematite, and it is believed that they have formed in part under arid conditions. Salt and gypsum are precipitated under arid conditions in bodies of water that have been cut off from the sea or in embayments that extend landward from the sea where the water flows into the embayment and is evaporated. This

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[1] Common salt, $NaCl$; gypsum, $CaSO_4.2H_2O$; anhydrite, $CaSO_4$.

process is illustrated in the gulf of Karabugas on the east side of the Caspian Sea.

The Karabugas gulf is only about 50 feet deep, and it covers about 7,500 square miles. It lies in a semidesert region and is partly separated from the Caspian Sea by sand spits, which form a barrier that prevents free circulation between the sea and the gulf. The waters of the Caspian Sea contain about 1.3 per cent mineral matter. Evaporation removes the water from the surface as fast as it is being brought in through the restricted connection with the sea. The water of the gulf becomes more highly concentrated than that of the Caspian Sea, and the less soluble salts are precipitated. Such marginal lagoons or gulfs are not uncommon; and in areas of great evaporation salt deposits may form in them.

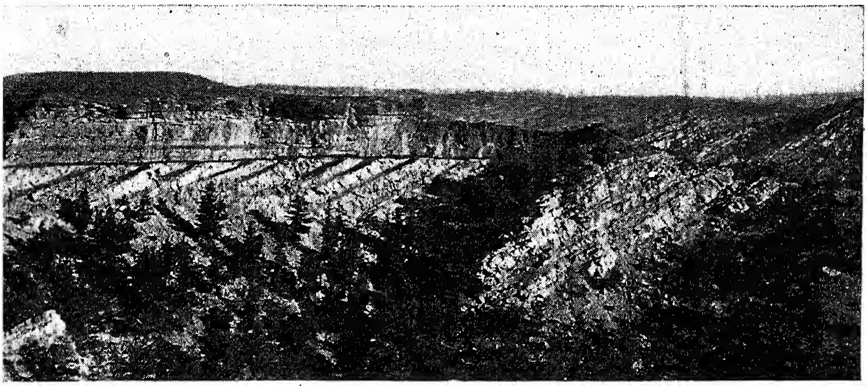
Salt water and brine often are encountered in deep drilling, and crystals of salt may occur in shales or in other sediments below the surface. The brines of eastern Ohio and adjacent regions contain bromides in addition to common salt. These brines may be sea water trapped in the sediments at the time of their formation, but some of them are more concentrated than sea water and probably have been changed greatly since they formed a part of the sea.

Bedded Iron Ore.—Iron in various forms is an abundant substance, and on weathering it may be dissolved and carried in solution to fresh-water bogs or to the sea where it is deposited by chemical or biochemical reactions. In certain beds it has formed in large amounts. These by surface enrichment may become iron ore.

Sedimentary Rocks of Lesser Importance

Adobe—argillaceous soil used for making sun-dried bricks.
Bone bed—a rock, commonly limestone, containing numerous bones, usually fishbones.
Buhrstone—porous cherty or siliceous rock formerly used for millstones.
Caliche—soil cemented by calcareous, nitrogenous, or other salts that rose and were deposited by evaporation at the surface.
Chalcedony—cryptocrystalline silica, probably a precipitate from a colloidal solution.
Catlinite—a red-clay "slate" containing abundant diaspore; carves easily and was used by Indians for making pipes. Pipestone.
Fire clay—a tough clay usually found underlying a coal seam. Makes refractory bricks.
Fuller's earth—fine earthy clay with low plasticity.
Guano—phosphatic materials formed from the excrement of animals.
Gumbo—clayey soils which become sticky mud when wet.
Ironstone—a rock containing iron, commonly with clay or sand.
Jasper—a siliceous red or variously colored rock resembling chalcedony.

Gradation of Sediments.—Sediments may grade laterally or vertically into other kinds of sediments forming intermediate or mixed types. On the usual gently sloping coastal portion of the continental shelf the coarser gravels are deposited near shore and are succeeded gradually by finer ebbles, by sand, by mud, then by calcareous mud and by ooze. Wave ction and shore currents, however, may disturb the regularity of this radation seaward and hence interfere with the regular order of sediments hat produce the sedimentary rocks. But the calcareous muds formed rom shell deposits and their fragments do not form limestones unless he quantity of land-derived sediment accumulating with them is small. imestones usually are formed in shallow water where lime-secreting



IG. 288*A*.—Photograph of an unconformity between the Laramie sandstone and the Wasatch conglomerate. (*Photograph by Fisher, U. S. Geol. Survey.*)

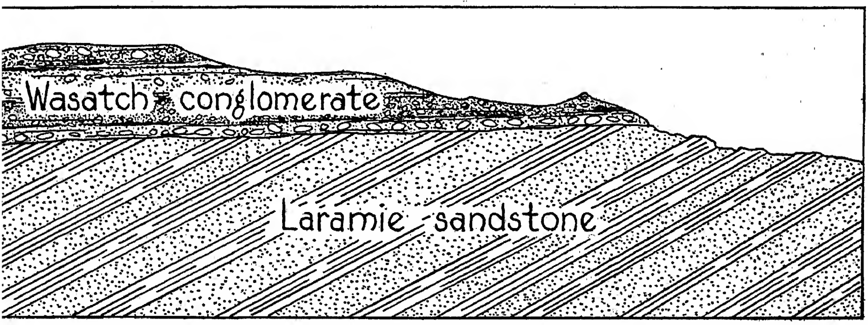


FIG. 288*B*.—Sketch of the unconformity shown in Fig. 288*A*.

fe is abundant and off of coasts that are near enough to base level o that the streams are carrying very little clastic sediment to the sea. vertical section of the sediments forming off a coast thus may show radations from conglomerate to sandstone to shale and finally to mestone as the adjacent land is reduced and land-derived sediments fail.

Such gradation produces a continuous series, and one kind of rock is aid to lie on the other with conformity. If, however, there is an erosion urface between any two beds there is evidently a discontinuity in eposition. A region of deposition may be elevated or uplifted and thus onverted into one of non-deposition or of erosion, and after an interval the same region may be depressed so that sediments again are deposited on the old surface. If the beds below the erosion surface are tilted so that they form an angle with the beds lying on top of it, the contact is called an angular *unconformity* or *nonconformity* (Fig. 288); but i the beds above and below the erosion surface are parallel, the contac is called a *disconformity*. The actual erosion surface in either case may be even or uneven depending on whether or not it had been base-leveled

In the normal course of events the sea may gradually encroach upon the land as the process of base-leveling advances. The deposition

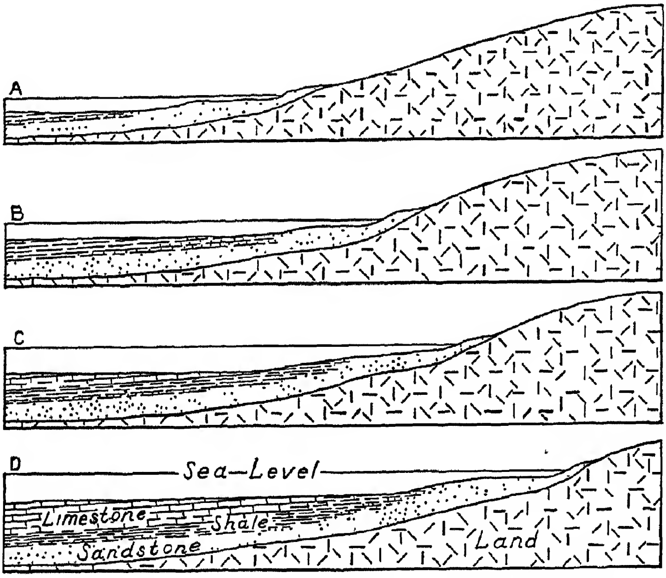


FIG. 289.—Diagrams showing progressive submergence of a land area. *A*, *B*, beach gravels and sands are deposited on the old land surface; *C*, muds that later form shale cover most of the sand; *D*, calcareous oozes that form limestone are deposited over most of the muds. This relationship is an *overlap*.

of the coarser sediments follows the retreating coast line, and therefore it takes place progressively farther to landward. This causes the newly deposited sediments to cover up the margin of the lately formed sediments and the basal beds to rest in turn on an erosion surface of continually changing age. The result is an unconformable contact known as an *overlap* (Fig. 289).

The reverse, however, takes place when the coast is rising and coarse sediments are laid down farther and farther seaward on top of the finer sediments. The sea is retreating, and each succeeding, or younger, division leaves a portion of the older one exposed to landward (Fig. 290). This type of contact between differing sediments has been called *offlap* (Fig. 290). The conditions which form offlap may be followed by those of overlap, and the top sandstone of emergence in the offlap may be followed directly by the bottom sandstone of submergence in an overlap, thus producing a compound, or double, series with the contact lying within the sandstone that represents both the emergence and the submergence.

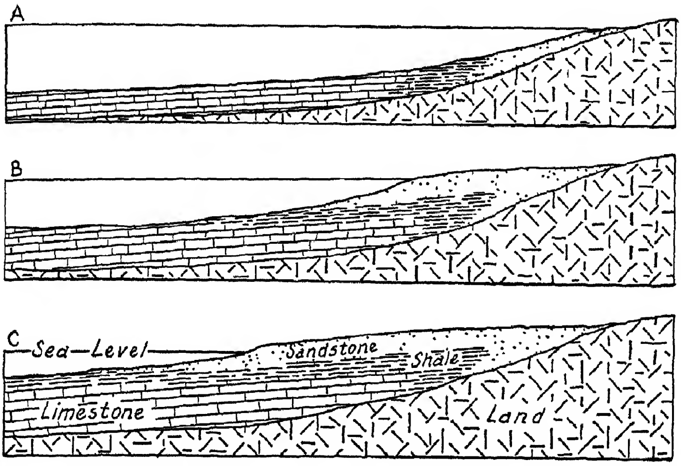


FIG. 290.—Diagram showing progressive emergence of a land area. *A*, beach sands are deposited on the old land surface; *B* and *C*, the land is lifted up slowly and, as the sea recedes toward the left, the beach sands are deposited over the fine offshore muds, and muds are deposited over calcareous oozes that later form limestone. This relationship is an *offlap*.

STRUCTURAL PECULIARITIES

Certain minor structural peculiarities may develop after sediments have been consolidated.

Concretions.—Concretions (Figs. 291, 292, 294III) are variously shaped masses or nodules of seemingly foreign material that often occur in sedimentary rocks. They range in size from less than an inch to several feet in diameter, and certain log-like cylindrical ones have lengths of 10 feet or more. Concretions usually differ in composition from the rocks in which they occur. Generally they are formed from one of the minor constituents of that rock. In limestone and chalk, concretions generally are of flint, chert, or pyrite; in shales, they are generally of calcite, chert, pyrite, or siderite, but barite and gypsum occur abundantly in certain localities. Concretions in sandstone generally are impure oxide of iron, pyrite, or calcite.

Concretions usually are spherical, lenticular, or discoidal, but more rarely they are cylindrical or irregular in shape. At many places the stratification of the beds in which they occur may be seen to thicken on entering the concretion, and those above are arched up, whereas those below are pressed down. Even cross-bedding may be found in concretions. At places fossils occur partly within the concretion and partly in the surrounding rock, or the center of the concretion may be one of the characteristic fossils of the formation in which it occurs. At some places concretions are abundantly fossiliferous, although the surrounding rock may be nearly destitute of them.

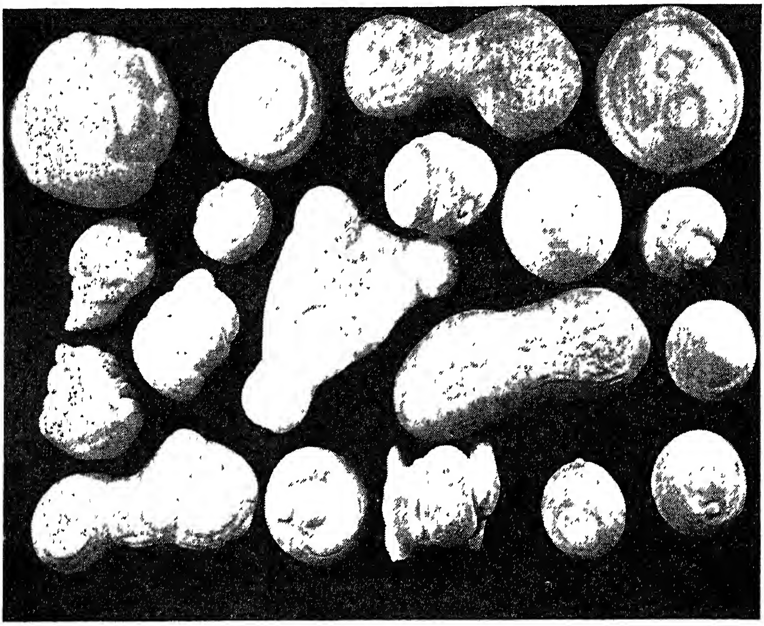


FIG. 291.—Concretions from clay beds.

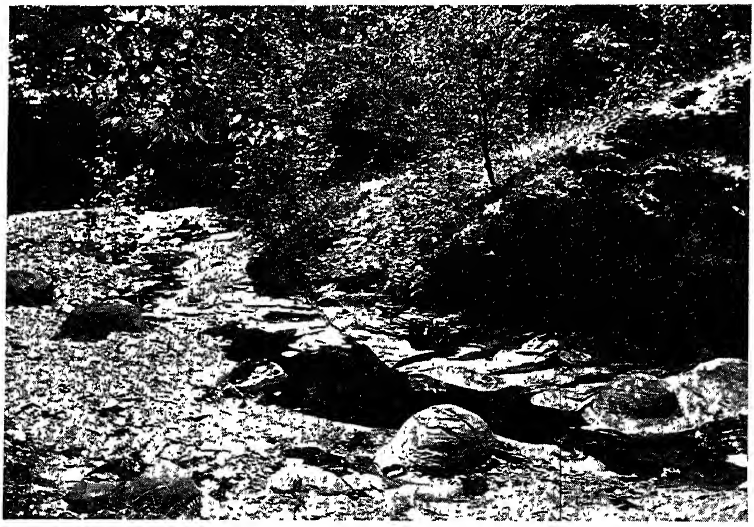


FIG. 292.—Spherical concretions weathering out of the Ohio shale near Worthington, Ohio

Concretions commonly show a concentric structure as if made up of a series of concentric spheres each fitting within the next larger. Those common in the black shales of Ohio are relatively smooth on the outside and separate easily from the enclosing shale. Many of them, in certain regions, are coated with a layer of pyrite, and rusty ones look like old cannon balls. At Kettle Point, Ontario, the large spherical concretions occurring in the Huron shale are composed of long, slender crystals of brown carbonate which radiate from the center. The concentric shells of some concretions differ in composition so much that weathering affects them differently. It thus happens that a shell, inside the surface layer, may disintegrate and leave the center loose. Such partly weathered concretions have been mistaken for fossil peaches or walnuts and are often referred to as rattle stones. It is evident that concretions were formed in place and since the deposition of the beds inclosing them, that they have grown up gradually from the center, and that they have partly displaced and partly incorporated the rock which originally occupied the places where they are found.

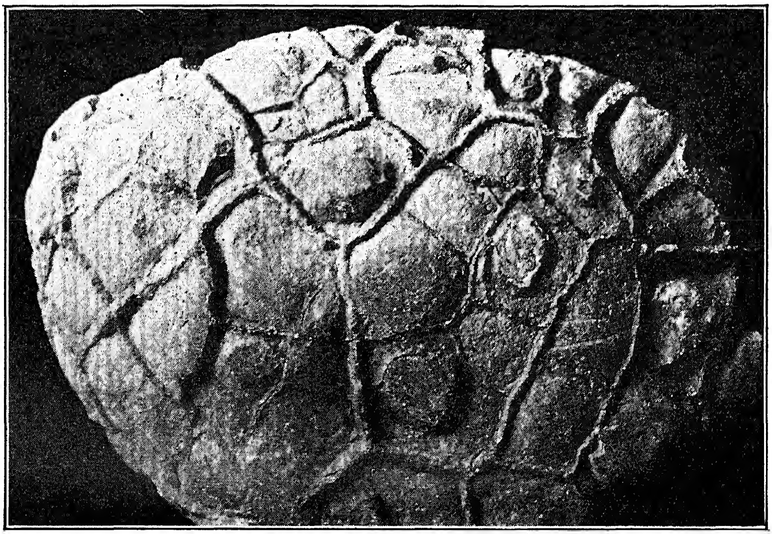


FIG. 293.—A septarian concretion.

The material of which the concretion is formed was probably once disseminated in the host, or inclosing rock, from which it was dissolved by ground water and carried to the place where the concretion is now found, often to be precipitated around some nucleus such as a fossil. Once started around the nucleus, precipitation continued, and the concretion increased in size.

Some concretions have been cracked or broken by jointing, and the cracks have been filled, vein-like, by material differing slightly in composition from that of the concretion. Such concretions are *septaria* (Fig. 293). Since usually they are disk-like in shape and show an irregular pattern resembling slightly the pattern on the back of a turtle, they have sometimes been mistaken for fossil turtles. Concretions are so abundant in certain limestones and in other sedimentary rocks that the whole rock is referred to as a concretionary mass. *Oölitic* limestone is made up almost entirely of small shot-like bodies crowded into a solid mass. The individual spherules are composed of concentric shells of calcite about some minute grain, such as a grain of sand or a fragment of a shell, and thus resemble very small concretions. When the spherules are about the size of peas, the rock is called a "pisolite." Oölites are found among the rocks of all ages, and in many the evidence suggests that the small spherules were formed during the deposition of the material that later was consolidated. Oölitic sands are now forming on the shores of Great Salt Lake and off some of the coral islands of the Pacific.

Geodes.—Geodes are cavities partially filled with crystals. The filling grows inward from the surface of the cavity, and the crystals in the hollow space commonly point toward the center. Most of them are composed of quartz or calcite, but some are composed of other material that was deposited from solution by ground water. They are common in all kinds of rocks.

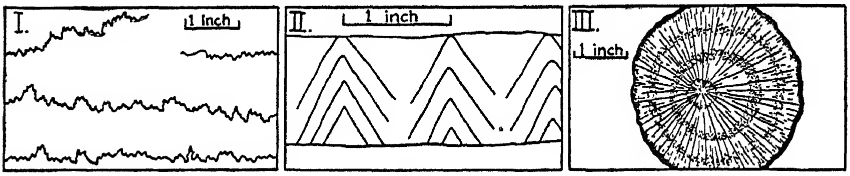


FIG. 294.—*I*, stylolites; *II*, cone-in-cone concretion; *III*, section of spherical concretion.

Cone-in-cone structure (Fig. 294, *II*) consists of a series of nesting concentric cones from a fraction of an inch to four or five inches high with bases somewhat smaller. These cones usually have wrinkled, fluted, or striated surfaces, but some may be relatively smooth, and in certain ones the conical surfaces are polished. The apices may be unusually sharp, and the bases flaring. The cones may occur singly or in various combinations with adjacent individuals. They frequently form a double series with the bases of the two sets in opposite directions. This structure is common in lenticular calcareous beds often in association with concretions in shales. Cone-in-cone is regarded as a secondary feature of layers containing fibrous calcite and probably the result of fracture aided by the chemical action of ground water, although compacting or pressure also may be a factor.

Stylolites (Fig. 294I) are striated or slickensided columnar and variously shaped projections of rock that form an interlocking and interpenetrating series along partings in carbonaceous rocks such as limestones. The projections vary in length from a fraction of an inch to a foot or more and are equally variable in width. Such sutured contacts may bind the two parts together so firmly that the rock will break as readily elsewhere as along the original parting plane. Stylolites result from different amounts of solution along a bedding plane or crevice, and their formation is promoted by the increased effectiveness of solution under different pressures at the points of contact on the two rock surfaces.

Color.—Sedimentary rocks may be white, neutral, or highly colored. The color generally is due to the presence of carbonaceous matter or to various iron compounds, chiefly the oxides. The carbonaceous matter usually occurs as finely divided organic material mingled with the sediment, but the iron compounds may occur as finely divided sediment, as cementing material, as coating for the individual grains of the sediment, or they may be within the mineral grains of the sediment.

The color of a sedimentary bed is characteristic only over limited areas and may differ with the degree of weathering to which it has been subjected. Thus many limestones having a wide range of colors may weather to buff, and the residual clays resulting from their solution may be red. Gray or black shales may weather to a red clay. The red iron oxide which colors the clay may be the oxidation product of pyrite, disseminated through limestone or shale so finely that its presence is scarcely detected in the fresh rock.

Fossils and Their Significance.—Fossils are the remains or traces of plants or animals preserved in the rocks (Fig. 287). Commonly they are the hard parts of plants or animals that are preserved. Since the plants as a whole have fewer hard parts suitable for preservation than animals, the plants are not so well represented by fossils as are animals. However, some very fine plant fossils do occur as impressions of leaves or stems in mud or sand deposits that later became shales or sandstones. Others have had their woody fibers gradually infiltrated with and filled in by silica, producing the fine specimens of silicified wood such as are found in the petrified forests of Arizona. Still other plants are preserved as coal, which is fossil vegetable matter. Cell structure may still be seen in thin sections of coal.

The bones, teeth, shells, and general skeletal matter of animals are more likely to be preserved; but the tracks, trails, burrows, or impressions of animals also may form fossils. In some cases the entire animals have been preserved, thus constituting unique fossils of great value. The best known of these are the fossil insects in the Baltic ambers and the woolly elephants frozen in the gravels of Siberia.

Such remnants of ancient life show its development through the long ages of earth history. Not only are the more primitive forms of life found to characterize the earlier periods in earth history, but the developmental changes of modern forms are recorded in the fossils that may be found in succeeding time intervals of that history. Thus the modern single-toed horse preserving the splint bones of two additional toes on each foot may be traced back to an earlier horse which had three functional toes, and that one back to a still earlier one which had five toes.

Age Relations.—Since fossils show the development from the more primitive to the more complex forms of life, it follows that a fossil-bearing rock is dated by the character of the life existing during the deposition of the rock that contains it. Thus geologists have come to recognize certain fossil forms as indices or guides to certain formations. Some of these are called horizon markers, because they are found only within certain horizons, or groups, of beds. If, then, the fossils occurring in the sedimentary rocks of two widely separated areas are alike, it follows that these sediments were forming at the same time; hence the rocks are of the same age. The sediments that were deposited continuously under similar conditions constitute a *formation,* and in an undisturbed series of such beds it is evident that the older beds lie at the bottom and the younger at the top. Formations therefore succeed each other in the order of their ages. The fossils likewise succeed each other in the order of life development. The fossils of one formation differ, therefore, from those of another, and a formation may be traced over wide areas, even through changing types of sediments, by the fossil fauna or flora which it contains.

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CHAPTER XI

VULCANISM

Vulcanism includes all phenomena that are connected with molten rock matter and its movements. The interior of the earth is very hot, and at places the heat is sufficiently great to melt solid rocks. The molten material rises toward the surface along rifts and fissures, dissolving and fusing its way. Where the molten matter solidifies before it reaches the earth's surface, it forms an intruding or intrusive rock. If it is thrown out or poured out upon the surface it forms an extrusive rock.

FIG. 295.—Mayon, a volcano in the Philippine Islands 7,616 feet high. For several centuries its vapors glowed at night, but the volcano is now quiet. (*After Chester H. Reeds, American Museum of National History.*)

Volcanoes.—A volcano[1] is an opening in the earth's surface through which hot rocks are thrown out. The rocks may be expelled in a fluid state as lavas or as solid rock fragments. Nearly always steam and other gases are ejected with the rocks and lavas. If solid matter or lavas accumulate around the opening, they build up a cone which increases in size and becomes a hill or mountain. A cone so constructed also is called a volcano, although it is a result of vulcanism and not a part of the volcanic mechanism.

Volcanoes vary in size from small conical hills to some of the loftiest mountains on the earth's surface. The Hawaiian Islands are volcanoes

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[1] The name volcano was first applied to Mount Etna in Sicily and to some of the Lipari Islands north of Sicily. It is derived from Vulcanus, the Roman god of fire, who was supposed to dwell in the volcano.

that reach a height of nearly 14,000 feet above sea level, and they are built on the floor of the Pacific Ocean where the sea is from 14,000 to 18,000 feet deep. Some of the highest peaks in the Andes are volcanoes. In the Cascade Range of western United States, Mount Baker, Mount Rainier, Mount Adams, Mount Hood, and Mount Shasta are volcanoes which recently have become extinct.

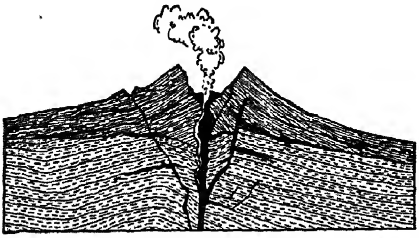

FIG. 296.—Geologic cross section showing structure of a strato-volcano. (*In part based on a figure by Heim.*)

Shapes of Volcanoes.—A hill or cone commonly is formed (Fig. 295) where volcanic material is thrown out. At the top of the cone there is generally a crater from which the volcanic matter issues. The slope of the cone usually is as steep as the rock material will lie. If the material thrown out consists of large solid fragments, the cone will have steep walls (Figs. 295, 296); some of them are as steep as 30 or 40 degrees. If the material consists chiefly of finely broken rocks or dust, water and wind will carry it farther away from the vent, so that the volcano will have gentler slopes. Volcanoes that throw out large fragments and dust are of the explosive type and generally eject gases in considerable amounts. Volcanoes of the explosive type include Vesuvius, Krakatao, Mount Pelée, and many others. Because of their stratified structure they are called *strato-volcanoes.* Certain volcanoes throw out little solid matter, their chief products being molten lavas. Gases are present; they do not accumulate under great pressure, however, but issue quietly. In such a volcano the lavas generally issue at high temperatures and are therefore more fluid. The basic lavas are more fluid than the acidic lavas and form more gentle slopes. Such cones are called *shield-volcanoes.* The volcanoes of Hawaii are well-known examples of this type (Fig. 297).

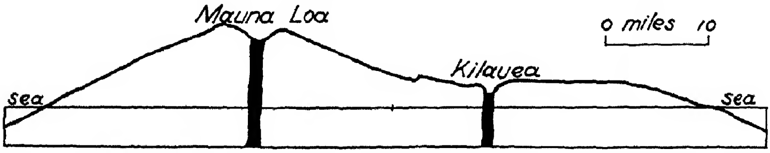

FIG. 297.—A cross section of Hawaii Island, showing the volcanoes Mauna Loa and Kilauea. Such volcanoes built up chiefly of lava flows are known as shield-volcanoes.

Craters.—The top of a volcano generally is marked by a pit, or crater. This is usually funnel shaped and represents the vent through which material is ejected, widened near the top by explosions and by the sliding back of the volcanic matter of the rim or by the lava in the crater melting and dissolving the rock of the funnel. The crater may be widened to great size, and subsequently upon its floor commonly another cone is built up of material ejected from the vent. This also develops a funnel-shaped crater, and later a third cone may form within it. Thus cones may nest one within the other (Fig. 298).

Parasitic Cones.—Vents are opened on the flanks of volcanoes, and material is ejected from them. From these vents are built up subordinate cones with characteristic slopes and craters. These volcanoes which are developed on the sides of older volcanoes are *parasitic cones.* It is believed that the central vent had become choked with the solidification of lava and that molten matter rose in fissures radiating from the central vent. Since the openings developed on the flanks are lower than the central vents, lavas more readily may rise from them (Fig. 299). Parasitic cones are developed on the flanks of Etna, Vesuvius, and other volcanoes.

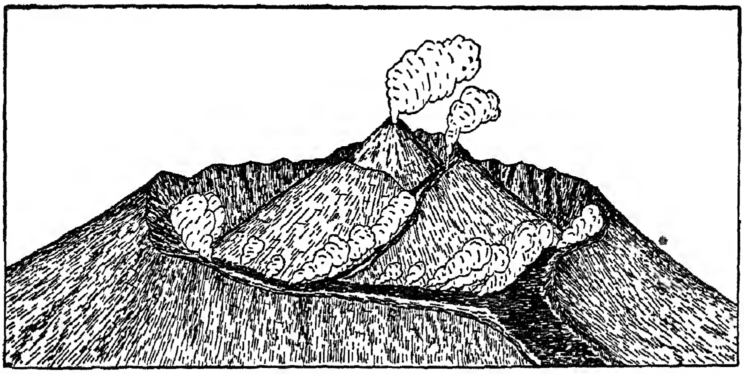

FIG. 298.—Sketch of a volcano showing a nest of craters. (*Based on a drawing of Vesuvius by Hamilton in* 1756.)

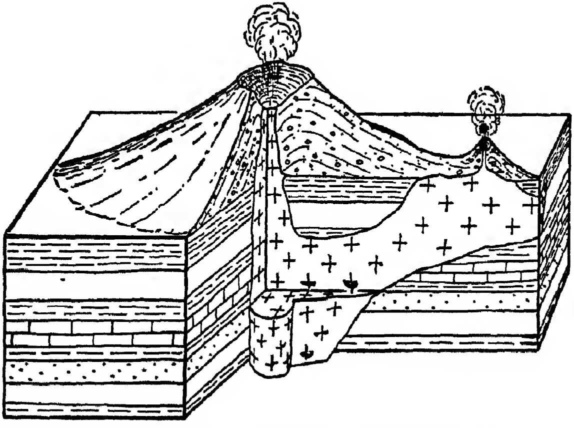

FIG. 299.—Ideal drawing of a volcano showing a radial dike and a parasitic cone. (*Based in part on a figure by E. Kayser.*)

Structures of Volcanoes.—Where volcanoes have been eroded and trenched by streams, their structures may be observed. The base rocks over which some of them are formed are nearly flat lying. Below others the rocks are arched near the volcano. In still others the beds dip toward the vents, where they have been dragged down into the craters.

The volcanic débris of which the cone is constructed is crudely stratified. The volcanic fragments and finer ash and dust lie in layers that are interbedded with flows and cut by dikes. The volcanic beds usually dip in the direction of the slope of the cone except near the crater, where at places they dip inward (Fig. 296). This figure shows a foundation somewhat disturbed by movement. The cone is made up of ash beds interbedded with lava flows. These and the basement rocks are intruded by dikes that extend outward from the central feeding channel. As soon as volcanoes have formed, erosion begins to wear them down. Lakes are formed in the craters. A crater is enlarged by running water, and often it is open on one side where a stream drains the lake. This enlarged crater has been called a *caldron*, or *caldera*.[1] Many caldrons have been enlarged by later explosions.

Eruptions.—The eruption of a volcano often is preceded by earthquakes and by loud rumblings, like thunder. These may continue during the eruption. During the eruption of Tambora on Sumbawa island, Netherlands East Indies, in 1815, rumblings were heard over an area with a radius of about 1,000 miles. During this eruption, which is the greatest one recorded by man, about 38 cubic miles of material were thrown out. The rumblings probably are due to the movements of gases and lavas that are held in under great pressure. Preceding the eruptions fissures often are opened, and lakes are drained. Hot springs appear at places, and flowing springs are dried up. Cracks open, and gases issue from them.

The materials erupted from volcanoes include solids, liquids, and gases. The fragments include lavas that have hardened in the throat of the volcano and rocks of the basement through which the volcanic channel passes. Liquid rock or lava may be thrown out along with solid material. Some of it solidifies in the air, forming "bombs."

Steam and other gases issue in large amounts. These gases, in the main, seem to rise upward with the lava, although some may form by reactions in the craters. Steam rises with the lava, but some of it forms when water flows over the hot lavas. Some of the most violent eruptions known are those near sea level, where it appears probable that water has penetrated fissures to the hot rocks and has been converted into steam. An example is the eruption of Krakatao, which took place in 1883. A small mountain was blown away, and below it a cavity

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[1] The name *caldera* has been applied also to wide valleys heading into volcanic areas where probably no crater existed originally.

1,000 feet deep was blown out in the sea bottom. A remarkable eruption took place in 1888 at the volcano Bandai-San, Japan (Fig. 300). The volcano had been dormant for more than 1,000 years. The top of it was blown off, the material removed being a mass of andesite more than a mile long and about 2,000 feet thick. No lava was ejected, and probably no volcanic gases rose. It is believed that the ground water in the rock became heated by the approach of molten matter or possibly by rise of gases and that pressures were accumulated that were relieved by the blowing off of the summit of the mountain.

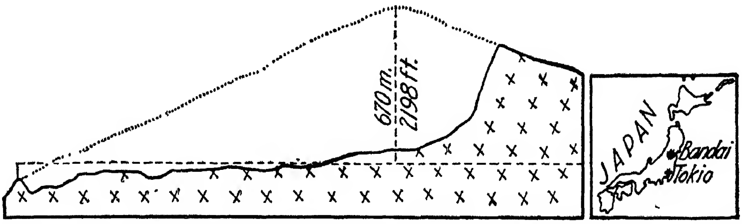

Fig. 300.—Cross section of Bandai-San, Japan. The dotted line shows the surface of the hill before the great explosion of 1888. The solid line shows the present surface. (*After Sekiya and Kikuchi.*)

Amount of Material Ejected from Certain Volcanoes
(Data from Sapper, Kayser, and others)

Name of volcano	Time of activity	Eruptive material	Amount of ejected material	Remarks
Bogosloff Island, Alaska	Formed 1796	Lava, ashes	....	New island formed near Bogosloff, 1883, which disappeared during 1907
Kluchewsk, Siberia	Still active	Lava, ashes	Eruption 1829 yielded 3⅔ cu. km. lava	East coast of Siberia
Bandai-San, Japan	1888	Ashes	More than 1 cu. km.	Explosive eruption; part of mountain blown away; no lava
Fujiyama, Japan	Still active	Lava, ashes	Uncertain	Highly symmetrical outline
Taal, Philippine Islands	1911	Ashes, gases	Unknown	
Tambora (Sumbawa Island)	1815	Ashes, slags, pumice	150 cu. km.	Greatest eruption in historic time
Bromo, Java	Still active	Ashes, lava	Uncertain	
Krakatao, Netherlands East Indies	Still active	Ashes, etc.	18 cu. km. in eruption	One of most violent eruptions of historic time
Mauna Loa, Hawaii	Still active	Lava	... .	
Tarawera, New Zealand	1886 last eruption	Ashes	1½ cu. km.	Great fissure formed during last great eruption

Many volcanoes erupt at intervals with explosive violence and are in mild activity at other times. After a period of quiescence a seal forms in the vent, and gases and lavas accumulate. These, under great pressure, are held in by the seal; but when the seal is broken by the pressure of the gases, a violent eruption follows.

Eruptions in Historic Time.—About 450 volcanoes have been observed in activity. There are also hundreds of others that have been affected very little by erosion and that must have been active in very recent time. A volcano in eruption is "active." One that has ceased to erupt is "dormant." If it is believed that the volcanic processes have subsided and that the volcano will not erupt again, it is said to be "extinct." It is difficult to distinguish between an extinct volcano and a dormant one, and many volcanoes supposed to be extinct have erupted with great violence. Mount Vesuvius was regarded as extinct in 79 A.D., when its first eruption in historic time occurred.

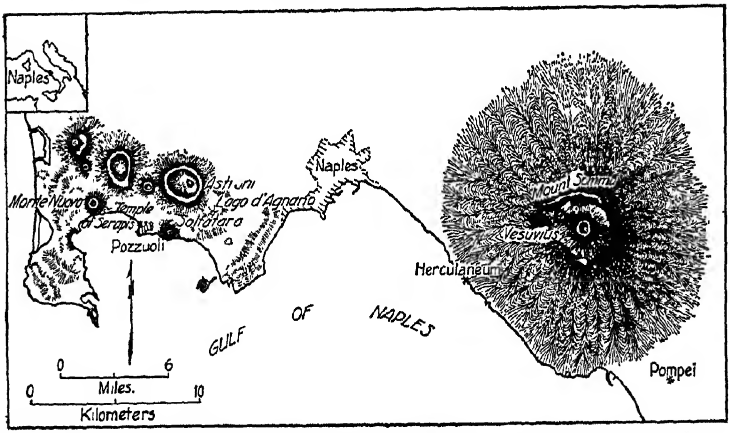

FIG. 301.—Ideal sketch of volcanic area near Naples. (*Based on a sketch by Shrope, with additions.*)

In the Phlegraean Fields, a densely populated area west of Naples (Fig. 301), in 1538 the eruption of Monte Nuovo took place. A new opening was formed, and in 3 days a cone was built up to a height of 500 feet. The eruption lasted a week, died down, and the volcano has not been active since.

Vesuvius.—Vesuvius, the best known volcano, is situated 7 miles southeast of Naples in a densely populated region. The height of the mountain varies but is about 4,000 feet above sea. Before 1906 it was 4,275 feet high, but that year it was reduced several hundred feet by explosions. A great ridge known as Mount Somma[1] half encircles the present active cone. This outer ridge partially incloses a huge crater formed in prehistoric time in which the present active cone is situated. In 79 A.D. the ancient crater had been quiet so long that trees were grow-

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[1] JOHNSTON-LEVIS, H. J., The Geology of Mount Somma and Vesuvius, *Quart. Jour.* Geol. Soc., vol. 40, p. 85, 1884.

ing in it. That year an explosion blew off a large part of the cone of Vesuvius, burying the cities Pompeii, Herculaneum, and Stabiae. Little or no lava was ejected, but much dust, ash, and steam issued, forming a pasty mud that flowed down the slopes and overwhelmed dwellings. Pompeii and Herculaneum were covered to depths of 25 to 50 feet. Gases were exhaled, and these either poisoned or suffocated the citizens. As shown by the positions of their remains, some perished while sitting at dining tables. Since 79 A.D. Vesuvius has erupted often. In 1631 a violent eruption attended by the ejection of lava, dust, and steam killed many people. Lava streams flowed to the sea, and dust was carried as far as Constantinople about 800 miles away.

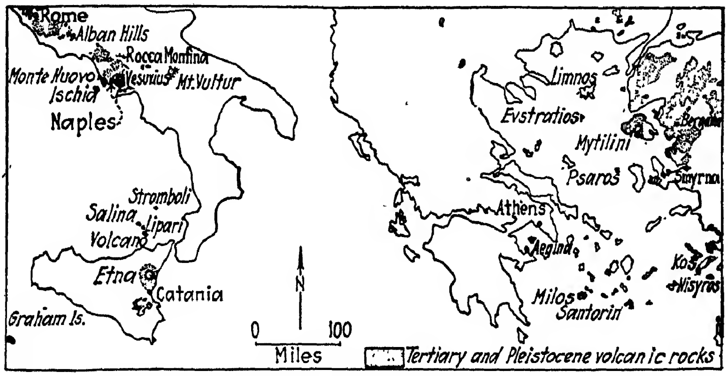

FIG. 302.—Map of part of Mediterranean volcanic area.

Phlegraean Fields.—West of the city of Naples (Fig. 302) is a volcanic area known as the Phlegraean fields, which extends along the coast for several miles. In this area volcanic craters are closely spaced, among them La Solfatara and Monte Nuovo. La Solfatara was in eruption in the twelfth century A.D., when a strong explosion took place. Sulphur fumes still rise from the crater. East of Pozzuoli (Fig. 301) there is a place called Cave of the Dogs, where fumes are emitted, and small animals have been smothered. In this region, near Pozzuoli, are situated the ruins of the temple of Serapis, which record the submergence and later the emergence of the area (page 342).

Etna.—Mount Etna (Fig. 302) on the east shore of Sicily rises to an elevation of about 10,758 feet and covers 460 square miles. The crater of the volcano is about 1,500 feet deep. Eruptions as early as the eighth century B.C. are recorded. A violent one in 1169 overwhelmed Catania, and many eruptions have occurred since. The temperature of a lava stream in 1892 was 1060°C. at the depth of one foot.

Lipari Islands.—Vulcano is an island of the Lipari group (Fig. 302). On the rim of its crater is fragmental material including great blocks weighing many tons. Acid vapors are exhaled.

Stromboli is one of the Lipari Islands and lies 35 miles north of the north coast of Sicily. It is 3,000 feet high, and its crater lies about 800 feet below the summit of the volcano. Steam issues from openings and forms a cloud which reflects the light of glowing lava. This light, often visible at night, served the navigators of ancient times, and Stromboli is called the "Lighthouse of the Mediterranean."

Krakatao.—One of the greatest eruptions recorded is that of Krakatao, a small island between Java and Sumatra. A great volcano existed in this area, and in prehistoric time its top was blown off, leaving a circular chain of islands around the crater. Subsequent eruptions built up small islands within the ring, one of which was Krakatao with its summit 2,623 feet above sea. In 1883 an eruption began throwing out pumice and dust, and on August 26 explosions began which lasted 3 days. Nearly all of the island was blown away, and a hole 1,000 feet below sea level was blasted out. Fragments were hurled 17 miles high, and in 15 days the dust from the volcano borne by air currents had encircled the earth. Sea waves 100 feet high were generated, and these reaching the shores of neighboring islands destroyed many towns.

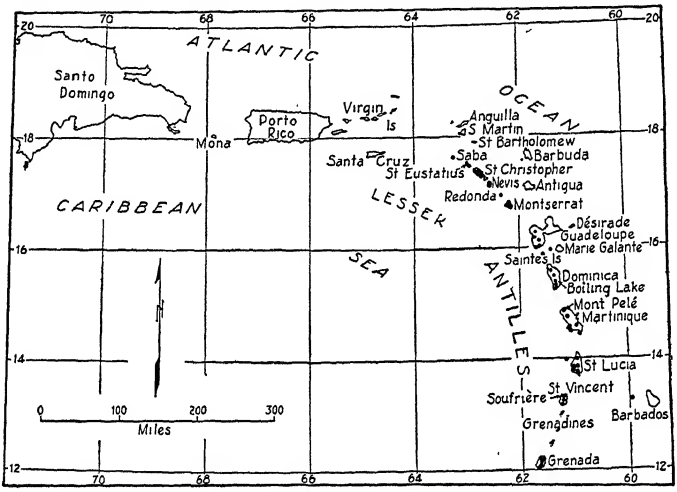

FIG. 303.—Map of volcanic area in Caribbean Sea. (*Volcanoes located chiefly from Sapper.*)

Mount Pelée.—Mount Pelée is situated at the north end of Martinique, an island of the West Indies (Fig. 303). On May 1, 1902, its crater blew out, belching dust and cinders. On May 8 great explosions occurred, and a great black cloud of hot gas and dust rolled from the mountain side like a hurricane and destroyed Saint Pierre, killing nearly 30,000 people. The cloud was so dense with dust that it seemed to act like a liquid. The destruction of life appeared to have resulted from the action of particles of hot dust on the membranes of the respiratory organs rather than from suffocation. In the course of the eruption a "spine" (Fig. 304) rose from the crater, its top reaching 1,000 feet above the level of the crater. The huge needle-like shaft of rock rose gradually above the crater, pushed up by pressure from below. Its sides were scratched and slickensided, and evidently it was made of solid or nearly solid material that had formed in the vent. The spine soon disintegrated, and in 1907 a mere stump remained, surrounded by broken fragments of the spine.

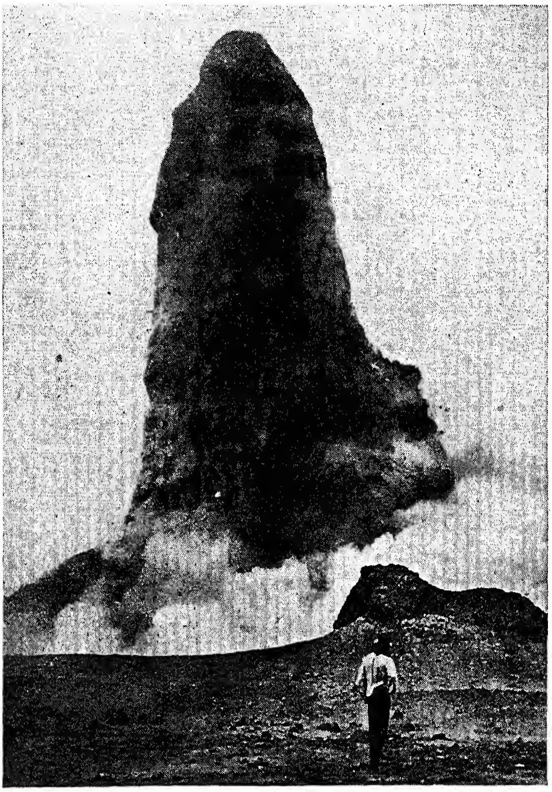

FIG. 304.—The rock spine of Mount Pelée which rose 1,000 feet above the crater floor. (*Courtesy American Museum of Natural History.*)

In recent years fiery clouds composed of lava particles and much gas rose with explosive violence from the crater.

Hawaii.—Hawaii, the largest island of the Hawaiian chain, is built up of volcanic matter, which rises high above the level of the sea. Mauna Loa in the southern part of the island has an elevation of 13,675 feet. It has a crater 2 miles wide and 1,000 feet deep. Kilauea, 4,050 feet high, is 20 miles away but is part of the great mountain mass.[1] The great difference in elevation of these two vents is assumed to indicate that they derive their lavas from independent sources[2] and probably

[1] Dana, J. D., Manual of Geology, p. 268, 1895.

[2] CHAMBERLIN, T. C., and R. D. SALISBURY, Geology, vol. 1, p. 605, 1905.

very deep ones. Mauna Loa, the "summit" crater, is not drained by the lower one. It is the more active in recent time and is often violent at periods when the lower crater is only mildly active. Numerous flows have been expelled extending downward from the summit crater, one reaching to near the sea (Figs. 305, 306). The crater of Kilauea is about 1½ miles in diameter. It is crusted over in the main, but in part of the crater the vent, Halemaumau, is active.

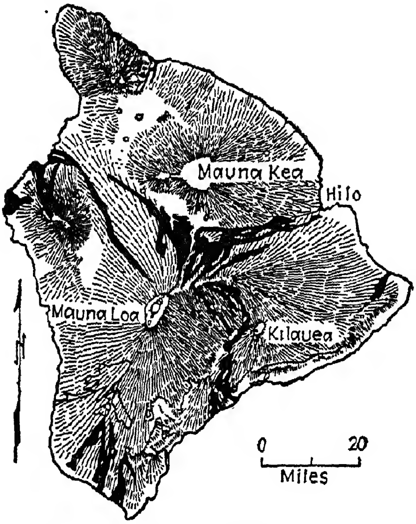

FIG. 305.—Map of Hawaii. Black areas represent recent lava flows. (*Redrawn from maps by Dana, Baldwin, Alexander, and others.*)

Mount Lassen.—In the western part of the United States there are many volcanoes that have been in eruption in late geologic times. Their cones and craters are still intact and are little affected by erosion. Several of these volcanoes are reported to have erupted in historic time, but these reports have been questioned. Mount Lassen in northern California is the only volcano in the United States except in Alaska that has had eruptions in the pres-

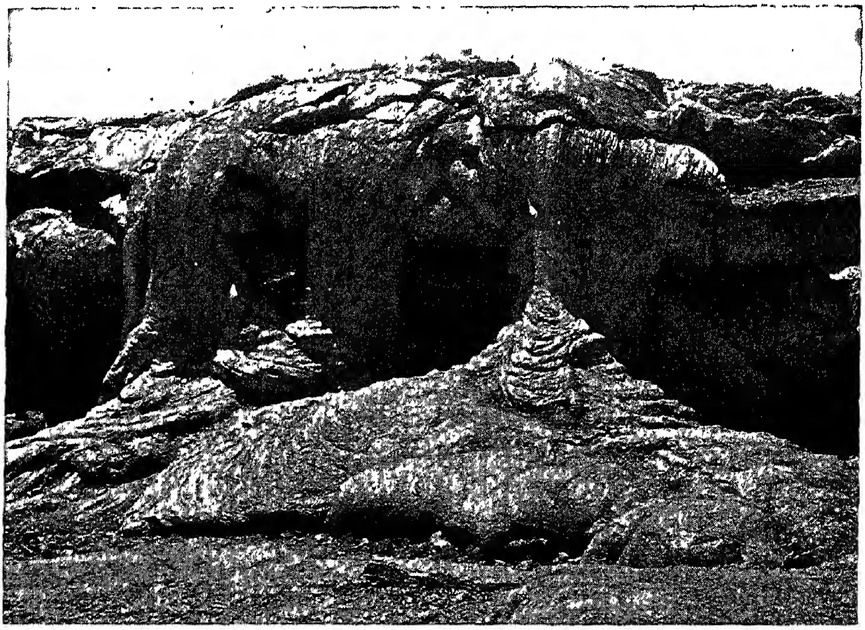

FIG. 306.—Ropy lava, Kilauea, Hawaii. (*Photograph by Mendenhall, U. S. Geol. Survey.*)

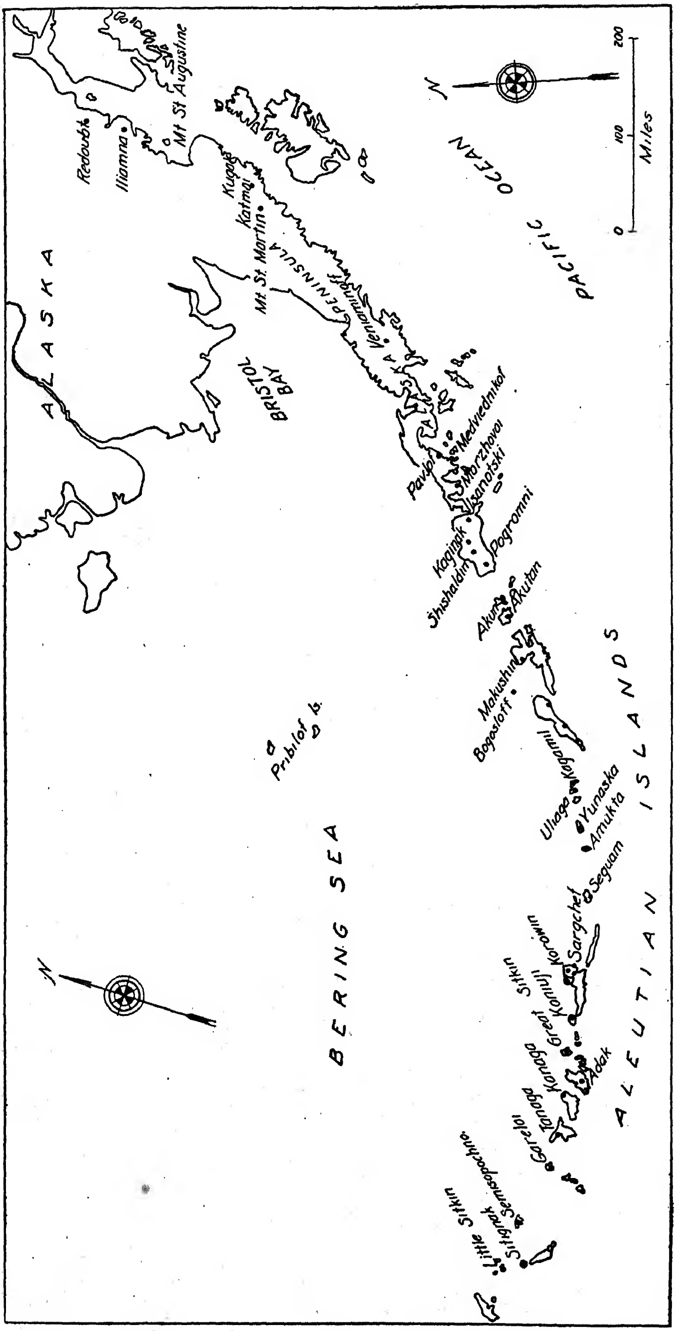

FIG. 307.—Map showing locations of volcanoes of Alaskan peninsula and Aleutian Islands.

ent century. These eruptions began in 1914, and there have been several since then. A small crater formed at the top of the mountain, and from it issued gases, ashes, and stones. Explosions took place in 1915. The eruptions of Katmai, Alaska (Fig. 307), are mentioned on a subsequent page.

Submarine Volcanoes.—Certain volcanoes erupt below sea level. About 80 such eruptions are recorded, and the history of some is well known. Some of the submarine eruptions were small. Evidently the supply of lava and hot material was insufficient to make great explosions when these came in contact with water.

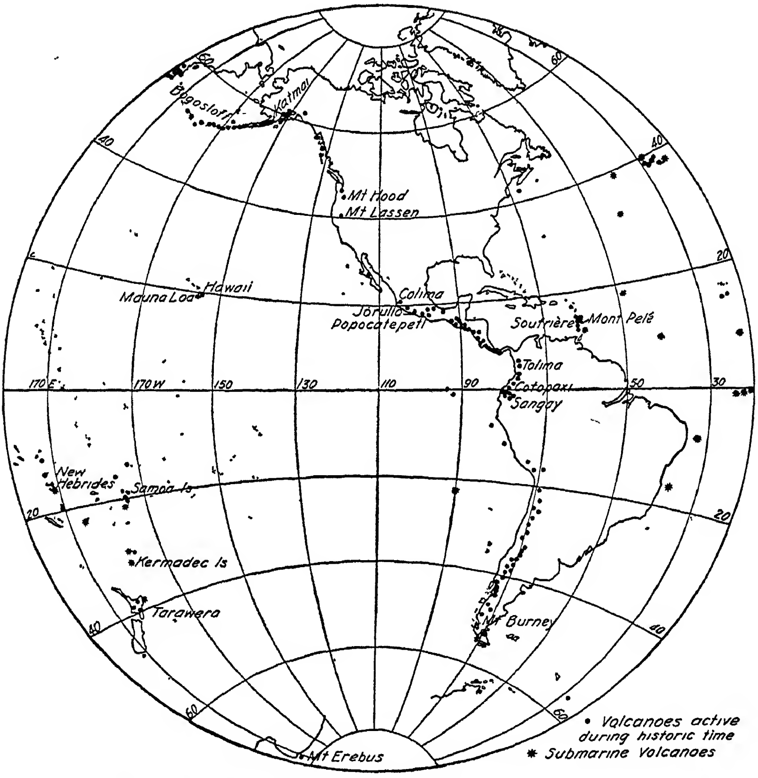

FIG. 308.—Map showing distribution of volcanoes in the western hemisphere.

Pantelleria, a small island which lies between Sicily and Cape Bon, Africa (Fig. 302), is wholly volcanic. On October 17, 1891, about 3 miles northwest of the island a submarine eruption took place. Red-hot bombs were hurled up, and an island 1,500 feet long and 9 feet high formed. This was soon washed away by waves. Between Pantelleria and Sicily in 1831 Graham Island rose from the sea, which had been 500 to 660 feet deep. The water boiled, and black clouds rose. Land appeared. Within two months a cone 200 feet high and about ½ mile in diameter was built up. Volcanic activity declined, and waves soon destroyed the island. Violent submarine eruptions repeatedly have been observed in the volcanic belt of the Aleutian Islands (Fig. 307).

Distribution of Volcanoes.—Wherever molten rock rises to the surface, volcanoes may be built up. They may form on mountains or plateaus, on low plains, and on the bottom of the sea. Volcanoes are widely distributed both chronologically and geographically. They were active at many periods in the geologic past, for their products are found in all of the great rock systems. In many series their products are very sparingly present, however, and in others they are concentrated. Evidences of them in the earliest known rocks are widespread. In very late geologic time also volcanic activity was widespread. In comparatively recent geologic time volcanoes existed at most places where they are active today and also in many other regions. In the intervening epochs there were periods of great igneous activity and other periods of relative quiescence over most of the earth's surface. Geo-

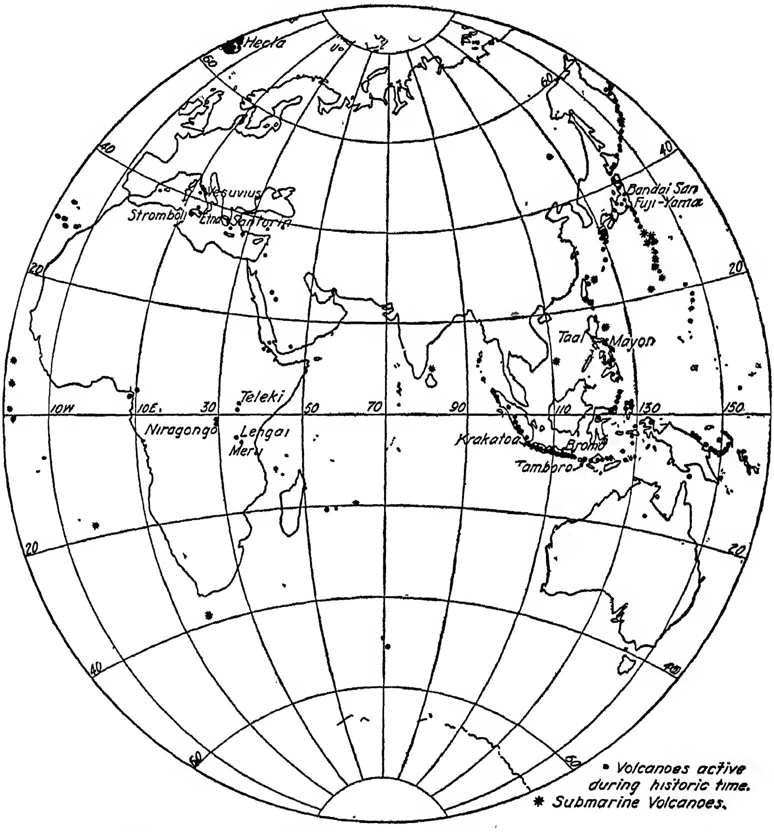

FIG. 309.—Map showing distribution of volcanoes in the eastern hemisphere.

graphically also, volcanoes are widely distributed, as is shown by Figs. 308 and 309. Many of the locations were obtained from a catalogue by Sapper.[1]

Locally volcanoes are grouped in belts, and it is believed that such are located along fractures or along fractured zones. This is suggested by Fig. 310. On the other hand, certain volcanoes seem to be independent of other volcanoes.

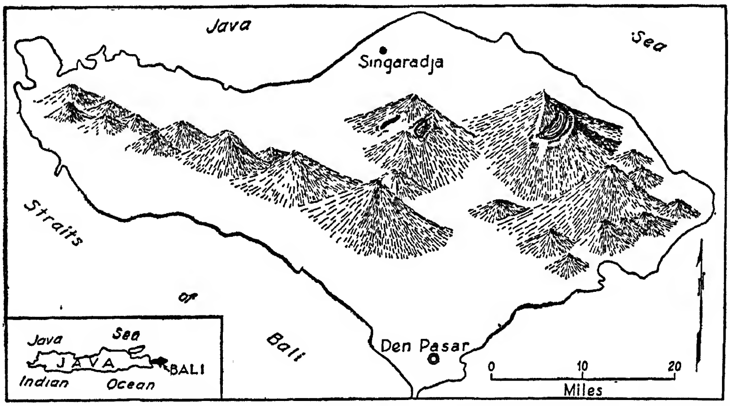

FIG. 310.—Sketch of Island of Bali with a group of volcanoes in line, suggesting their occurrence along a fissure.

The Pacific Ocean is essentially bordered by a volcanic belt which has been called the "circle of fire." Another belt extends westward from Baluchistan, through Asia Minor, the Mediterranean, and the Canary and Azores islands. In two areas volcanic activity is notably concentrated. One of these includes the West Indies, northern South

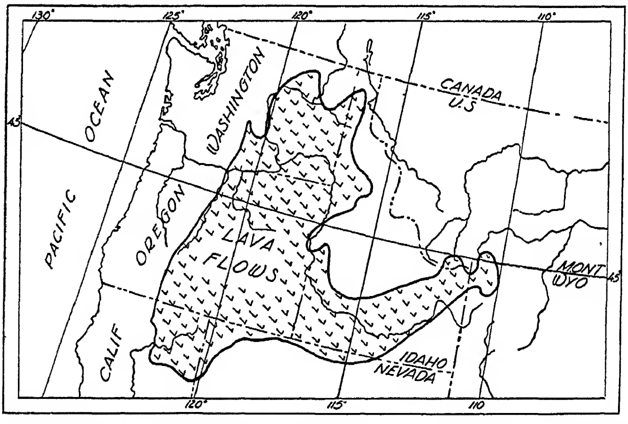

FIG. 311.—Map showing tertiary lava flows in northwestern United States.

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[1] SAPPER, K., Katalog der geschichtlichen Vulkanausbrüche, Schriften der Wissenschaftlichen Gesellschaft in Strassburg, Heft 27, 1917.

America, Central America, and southern Mexico. The other is on the opposite side of the Pacific Ocean and nearly east of the West Indies group. It includes Sumatra, Java, New Guinea, and the Philippine Islands. In addition to these volcanic areas around the Pacific Ocean, volcanoes are found at many places in the Pacific, where they are arranged in northwest-trending zones. A great belt of the Atlantic Ocean includes Spitzbergen, Jan Mayen Island, Iceland, Azores, Madeira, and the Canary and Cape Verde islands. A third belt extends from Palestine southward through Arabia, the Red Sea, Abyssinia, to East Africa. There are also many apparently isolated volcanoes in the Pacific Ocean. These areas include nearly all of the active and recently active volcanoes of the world.

Fissure Eruptions.—Lavas commonly rise along fissures. At many places volcanoes are arranged in lines which suggest a connection with fissures. In Iceland lavas have issued from fissures in recent times. At many places in the world there are great areas that are covered by flows of basalt or other lavas which, taken together, amount to hundreds or thousands of feet in thickness (Figs. 311, 312). The basalts were poured out in a very fluid state, so they covered valleys like floods of water. Single flows about 60 miles long are known. Certain regions with lava flows contain no large volcanic cones and few volcanic cones of any kind, and because of their absence it is commonly assumed that the lavas issued from fissures. H. S. Washington,[1] who studied many of these flows, suggests

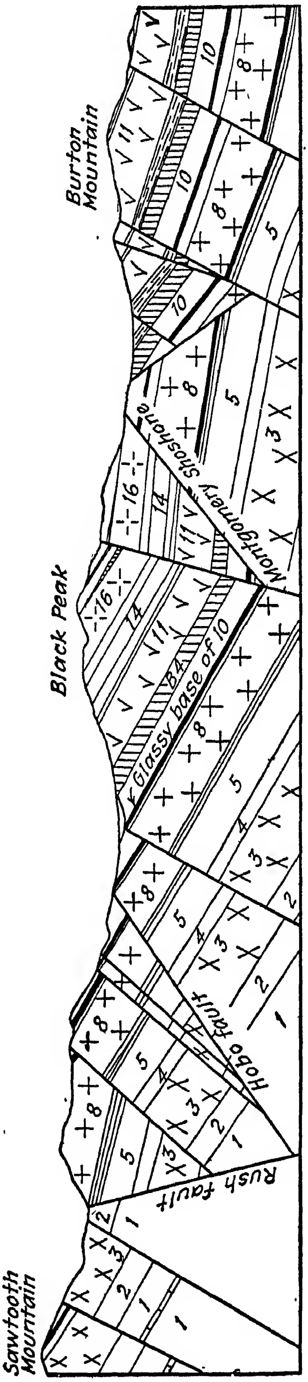

FIG. 312.—Section of Rhyolite district, Nevada. The numbers refer to different lava beds that are complexly faulted. The heavy black lines are glassy bases of rhyolite flows, and the vertically ruled beds are basalt flows that are interbedded with the rhyolite flows.

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[1] WASHINGTON, H. S., Deccan Traps and Other Plateau Basalts, *Bull.* Geol. Soc. America, vol. 33, pp. 765–804, 1922.

that they be called *plateau* flows, not because all the lavas are found high above sea level but because the word implies flatness or horizontality, which is characteristic of such flows.

Some of these basaltic flows are of very great extent. The Deccan flows of India cover an area of 200,000 square miles and probably average 2,000 feet in thickness. In the northwestern part of the United States, in Washington, Oregon, and Idaho, the Columbia River basalts cover about 225,000 square miles and have an average thickness of about 500 feet. Large areas of similar rocks are found also in northern Michigan and Wisconsin. Washington studied the rocks from all of these regions and has shown that they are characterized by high iron content, particularly by high ferrous iron, and for that reason were highly fluid lavas. On account of their lower viscosity, basic lavas move faster and farther than acidic ones. Some of them are known to have moved as fast as 10 or 12 miles an hour. In Hawaii a basic lava has flowed more than 40 miles, and one flowed in Iceland nearly 60 miles.

The low viscosity accounts for the absence of great volcanic cones and in part for the wide distribution of the flows. Some of these rocks are highly vesicular and must have contained gases, yet their fluidity was such that their eruption was attended by little violence.

Caves in Lava Flows.—Some lava flows show great caves that probably were formed by the flow of lava from below the solidified crust. Government Cave, 23 miles west of Flagstaff, Arizona, is a cavern in lava which flowed from the west side of San Francisco Mountain. The cave is about ¾ mile long and from 20 to 50 feet high. A thick flow of lava was poured out from a volcanic opening, and the upper part, being exposed to the air, cooled first. Later a low opening was provided, and the liquid lava flowed out from below the solid roof, leaving the cavity as it now is. Small stalactites are noted on the roof, which at places is 45 feet high.

Spatter Cones.—As a lava cools, a crust forms on its surface. The moving of the lava causes the crust to fracture. If gases have accumulated below the solid crust, the portion still liquid may be under pressure due to impounded gas. The lava and gas are expelled through the fissures, and the molten lava builds up a small "spatter" cone, or "driblet" cone. Such cones commonly are 15 or 20 feet high or less. They probably are not connected with the main volcanic vents but are merely features of the solidifying lava mass.

Mud Volcanoes.—Gas issuing at the surface of the earth may carry with it particles of sand and clay which are deposited at the vent. As the process continues, a cone is built up. With water the sand and clay form mud, which dries and hardens at the surface of the mound. Gas accumulates below the hardened surface until the pressure is sufficient to blow off the top of the cone, imitating on a small scale the eruption of a true volcano. Some of the mounds, or "volcanoes," are built to considerable heights. The famous Bog-Boga mud volcano in the Baku region near the Caspian Sea is more than 100 feet high. Many mud volcanoes are found in oil fields, and some of these are far removed from true volcanic areas. They are formed from the gas that escapes from gas-bearing strata. Other mud volcanoes are found in areas where steam, probably volcanic, escapes through mud.

Volcanic Gases.—Gas issues from certain volcanoes in large amounts. Steam is by far the most abundant. Some of it is formed by heating ground water and surface water that have come in contact with the hot products of the volcanoes. Much of it, however, is believed to be steam that was dissolved in the magmas. Chlorine and sulphur gases issue with steam from certain volcanoes, and these may be in part the products of heated sea water but probably in a large measure are of magmatic origin.

The gases arising from the Valley of Ten Thousand Smokes[1] a few miles northwest of Katmai, Alaska (Fig. 307), have been closely studied. An eruption in 1912 scattered ashes 6 inches deep as far as 150 miles away. In this valley steam issues from hundreds of vents in large amounts, and it is probable that a buried intrusive lies below the valley. Hydrochloric and hydrofluoric acids, sulphur, and boron compounds rise with the steam. Zies estimated that 1,250,000 tons of hydrochloric acid and 200,000 tons of hydrofluoric acid issue annually. The steam and chlorine compounds might be derived in part from sea water, but that is improbable, since they are associated with large amounts of hydrofluoric acid, and fluorine is very sparingly present in sea water. The steam which constitutes 99 per cent of the gas issuing from the vents is very hot, about 97° to 650°C. The steam and vapors carry with them many metals in appreciable amounts, among them iron, lead, zinc, molybdenum, copper, arsenic, antimony, tin, and silver, which were identified in the incrustations along fissures through which the gases rise. The gases issuing from a vent of Mount Pelée in 1902 were examined by Lacroix and were found to consist of water vapor, with hydrogen chloride, carbon dioxide, carbon monoxide, sulphur, methane, hydrogen, nitrogen, oxygen, and argon. These gases issued at a temperature of 400°C.

In certain eruptions volcanic gases other than steam seem to be essentially absent.

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[1] Griggs, R. F. T., The Katmai Region, Alaska, and the Great Eruption of 1912, *Jour. Geol.*, vol. 28, pp. 569–606, 1920.

Zies, E. G., The Valley of Ten Thousand Smokes, Nat. Geog. Soc., *Tech. Paper* 4, vol. 1, pp. 1–79, 1929.

GASES ISSUING FROM VOLCANOES AND FUMAROLES

Steam, H_2O
Oxygen, O
Nitrogen, N
Argon, A
Carbon dioxide, CO_2
Carbon monoxide, CO
Sulphur dioxide, SO_2
Hydrogen sulphide, H_2S
Hydrochloric acid, HCl
Hydrofluoric acid, HF
Ammonia, NH_3
Sulphuric acid, H_2SO_4

Fumaroles.—Fumaroles[1] are vents in the rocks of the earth's crust from which gases issue. They are common in regions of active volcanoes and also in areas of decadent vulcanism, where no active volcanoes now exist. Lavas, where deeply buried, cool very slowly and supply heat long after they are expelled. Many intrusives do not reach the surface but probably supply heat and gases to ground water. Aside from steam, which constitutes the chief issue of the fumaroles, the gases present include carbon dioxide, hydrogen sulphide, sulphur dioxide, chlorine, hydrochloric acid, hydrofluoric acid, and others. The steam of fumaroles is in part from ground waters, but some of the water probably is derived from the intrusives, as are some of the other gases. Solfataras[2] are fumaroles which give off sulphur gases. At places the hydrogen sulphide gases oxidize on exposure to the air and form sulphur which accumulates in considerable amounts, so that the rocks near the solfataras are worked commercially for sulphur.

The temperature of the steam that issues from certain fumaroles is well above the boiling point of water. Temperatures as high as 650° C. have been measured.

Poisonous Gases.—Hydrogen sulphide and carbon monoxide are deadly gases. Sulphur dioxide and carbon dioxide are suffocating gases, the latter collecting at low places. Where these gases issue from vents, animals are killed. In Poison Valley, Java, it is said that the bones of men and animals killed by gases have been found. In Death Gulch, Yellowstone Park, animals seeking warmth are killed in large numbers. Vents from which carbon dioxide issues are *mofettes.*

Volcanic Sulphur Deposits.—One of the best-known vents yielding sulphur compounds is La Solfatara, west of Naples (Fig. 301). The last eruption of this volcano occurred in 1198 A.D. Since then steam and sulphur compounds have issued. At places in Mexico, Japan, and elsewhere volcanic emanations are worked commercially for sulphur. This is formed by the action of the air on hydrogen sulphide vapors, as follows:

$$H_2S + O = H_2O + S$$

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[1] From the Latin, *fumariolum*, smokehole.

[2] Solfatara, Italian, *solfo*, sulphur.

Hot Springs.—Hot springs commonly are found in areas of late vulcanism, but they are not confined to such areas (page 72). The heat and probably some of the water of certain hot springs are of magmatic origin, but some hot springs are in areas remote from late igneous activity and probably have no generic relation with igneous rocks.

Utilization of Volcanic Gases.—In Tuscany, Italy, plants for generating electric power and for producing boric acid are erected. In this volcanic area there are steam jets and pools of hot water boiling by natural heat. Boreholes are drilled and cased with iron tubing. Steam issues at an average pressure of 2 atmospheres and at 100° to 190°C. At Larderello 150 metric tons of steam per hour is available from 135 boreholes. It is used in turbine engines to generate electric power. From the steam and the hot water ammonium carbonate, sodium carbonate, and boric acid are recovered. In the Coast Range of California, about 40 miles north of San Francisco, in an area of hot springs, wells have been drilled to obtain steam for generating power.

INTRUSIVE VULCANISM

Volcanoes are the surface expressions of igneous intrusives that extend downward to great depths (Fig. 313). One may not observe the operations of volcanic processes at depths, for the roots of the volcanoes are concealed, but at many places erosion has removed the rocks and exposed the lower portions of ancient volcanoes that were formed ages ago. There are no glassy lavas at the roots of the volcanoes, yet for every fine-grained or glassy surface flow rock there is a deep-seated granular rock type of similar chemical composition. The same magma solidifying at depth and forming a coarse-grained igneous rock, if spread out upon the surface, would have formed a glassy or fine-grained flow.

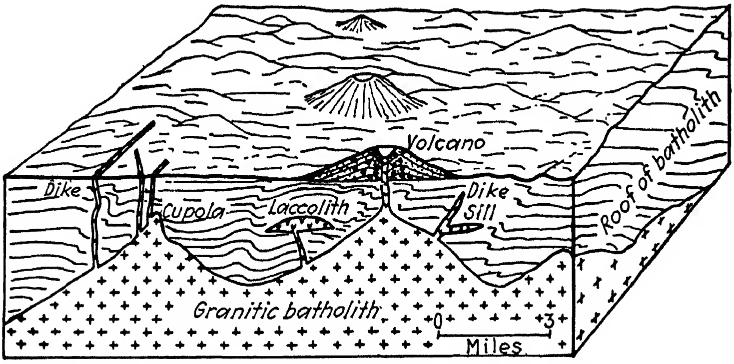

FIG. 313.—Diagram showing relations of intrusive and extrusive igneous rocks. Molten matter rose to form the body designated as the granitic batholith. Parts of the magma rose from the batholith to the surface and were poured forth as liquid or thrown out in solid state by volcanoes. Molten matter was thrust into fissures to form dikes, and where injected between beds, it gave rise to sills and lacoliths.

Igneous intrusives occur in many shapes and show various relations to the rocks which they invade.

Forms of Intrusives.—Dikes are intrusives rudely tabular[1] in outline that fill fissures near the earth's surface (Fig. 314). They vary from less than an inch to many feet in width. Some of them may be followed for a mile along the strike, and a few have been traced many miles. In certain regions dikes radiate from intrusive centers (Fig. 315).

Fig. 314.—Pegmatite dike with offshoots, Cornwall. (*Geological Survey and Museum, London.*)

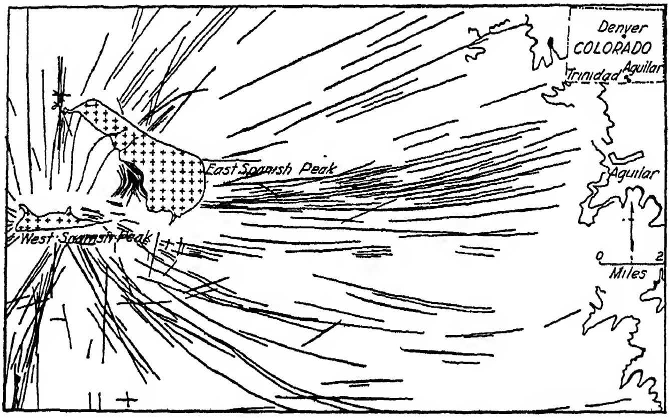

Fig. 315.—Map of Spanish Peaks area, Colorado, showing radial dikes extending from intruding stocks.

Sills also are intrusive tabular masses. They are essentially like dikes, but they lie parallel to or nearly parallel to the bedding planes

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[1] Tabular—like a tablet, long in two and short in one dimension.

of the rocks that they invade. The term "sill" is applied to thin, flat bodies, like the sill of a door, but inclined bodies parallel to the beds are sills also. Some of them are small, covering areas of a few acres. Others are very extensive. The great Whin Sill of northern England has an average thickness of about 160 feet and extends over an area of several thousand square miles. This sill intrudes limestone, is nearly flat lying, and in general is almost parallel to the beds it intrudes but cuts across them locally, as is shown by Fig. 316.

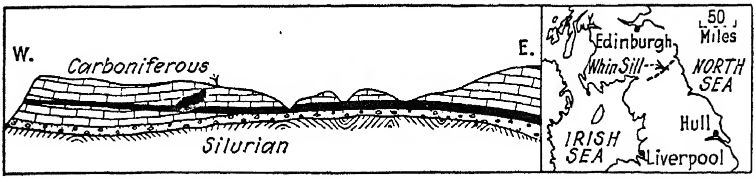

Fig. 316.—Diagram showing cross-cutting relations of part of the Great Whin Sill (black) in northern England. (*Section from Lake and Rastall, index map from Harker.*)

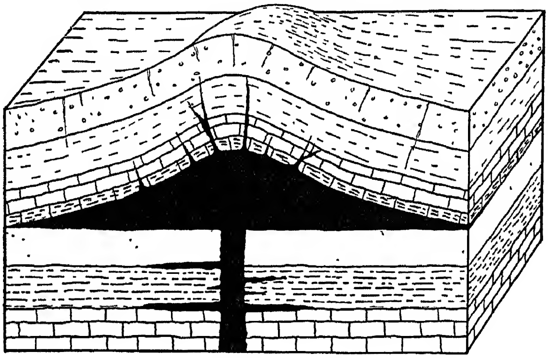

Fig. 317.—Ideal cross section of a laccolith. The igneous rock (black) intrudes the beds and arches them up.

If a sill lifts up the overlying beds or raises its cover arch-like, it is a *laccolith* (Fig. 317). A *bysmalith* is a laccolith faulted on its sides (Fig. 318). The common conception of the laccolith is that it is intruded into essentially flat-lying beds and that it has an essentially flat floor of the invaded rock. After a laccolith has formed, or possibly during its formation, the floor may sag down so that the sheet forms a basin. Such a

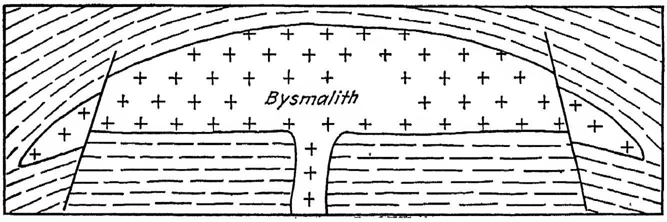

Fig. 318.—Cross section of a bysmalith—a laccolith faulted at the margins.

laccolith is called a *lopolith* (Fig. 319). A *neck* is a more or less cylindrical body through which lava is supposed to have risen to the surface. A *stock* is a larger, dome-like intrusive rudely circular in horizontal cross section.

There are certain great irregular masses of coarse-grained rocks formed at great depths, which are called *batholiths* (Fig. 313). These differ from the laccoliths in that none is known to have a floor. They extend downward to great but unknown depths, and nearly all of them, where their contacts have been observed, are found to broaden downward. It appears improbable that they broaden downward indefinitely, but nothing is known of their lowest extensions. They are the deepest seated and largest bodies of intrusive igneous rocks known and are believed to have been the feeding masses of laccoliths, dikes, sills, and volcanoes.

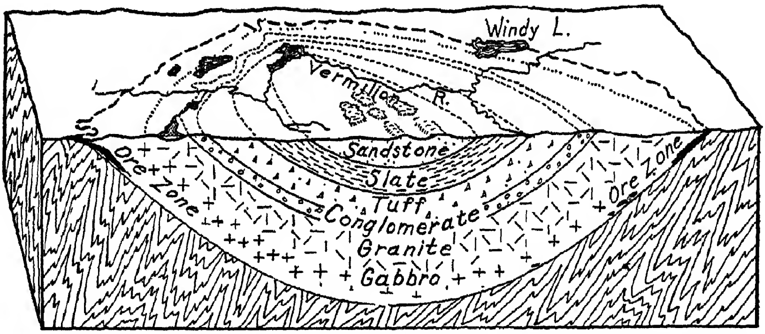

FIG. 319.—A lopolith. Cross section and plan of part of Sudbury nickel district, Ontario. (*After Coleman.*)

Laccoliths in general are relatively small bodies; some are less than a mile in diameter. Most of them are not more than a score of miles in diameter. A few larger ones are known. Batholiths, on the other hand, are very extensive. Most of them are aligned parallel to mountain ranges, except in very deeply eroded areas, where such alignment is not evident. Some are from 50 to 100 miles wide and are hundreds of miles long. Their roofs are irregular and undulating. The small, dome-like parts of their roofs that extend upward into the invaded rocks are *cupolas*, and the low sags of the invaded rocks that project or hang downward from the roof are *roof pendants* (Fig. 25). Most batholiths are composed essentially of granite and nearly related granitic rocks.

The magma rises toward the surface partly by pushing its way up along fractures and by opening fractures. In many batholiths, however, the rocks of the roofs above the intrusive rocks are not greatly disturbed and are not steeply tilted. The magma seems to rise also by melting or fusing its way up or by dissolving the overlying rock. Along with this process "magmatic stoping" also is carried on; that is, large blocks are loosened from the roof, spalled off by heating, and fall in the liquid mass, where probably they are dissolved in part or sink to great depths.

Jointing in Intrusive Rocks.—Igneous intrusives on cooling often develop characteristic cracks or joints. Experiments have shown that when a rock cools 1°C., it contracts about 0.000028 of its volume. A cooling of 500°C. would cause a shrinkage of about 0.014, or 1.4 per cent. The joints in many intrusives are arranged in definite systems (Fig. 320) depending upon the rate of cooling, the size and shape of the mass, and other features. All deep-seated rocks are jointed. The great bodies of granite and other deep-seated rocks commonly are cut by three or more sets of planes that divide them into irregular prisms. A common type is the columnar jointing, which is developed in many tabular intrusives, such as dikes and sills (Fig. 321). In these the joints lie normal to the cooling surfaces. In flat sills the columns are upright, and in vertical dikes they are horizontal. Near Spanish Peaks, Colorado (Fig. 315), some of the dikes may be followed for miles and form great ridges that are capped by the piles of horizontal columns. Viewed at a distance the dikes resemble huge piles of cordwood.

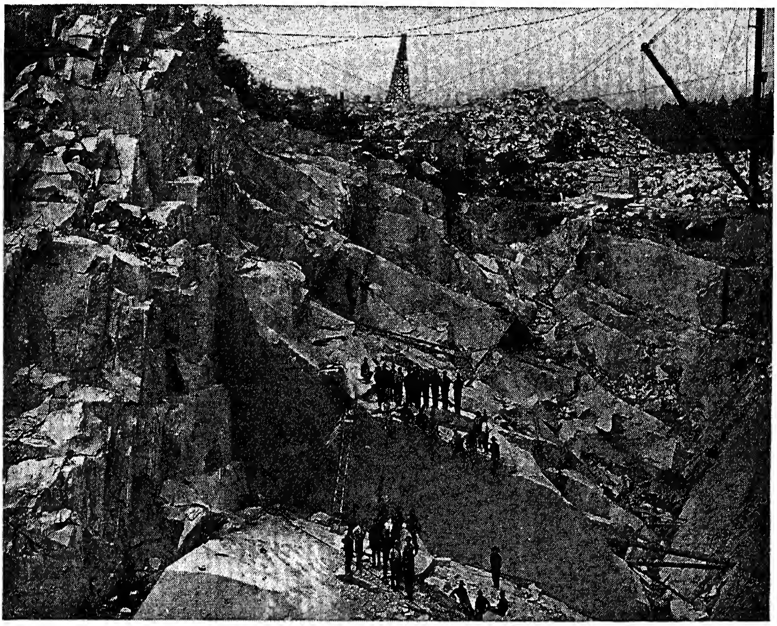

FIG. 320.—Jointing in granite, Barre, Vermont. (*Courtesy Oliver Bowles.*)

Age Relations of Intrusives.—Except in rare instances, igneous rocks do not contain relics that may be used to determine the age of the rocks as fossils are used to determine the age of a sedimentary bed. The relative age of an igneous rock frequently may be determined, however, by observing its relations with associated rocks. If a rock intrudes another rock, it is younger than the rock invaded (Fig. 322), and if it is covered by another flow or sedimentary bed that was laid down above it, it is older than the flow or bed. This is illustrated by Fig. 323, which

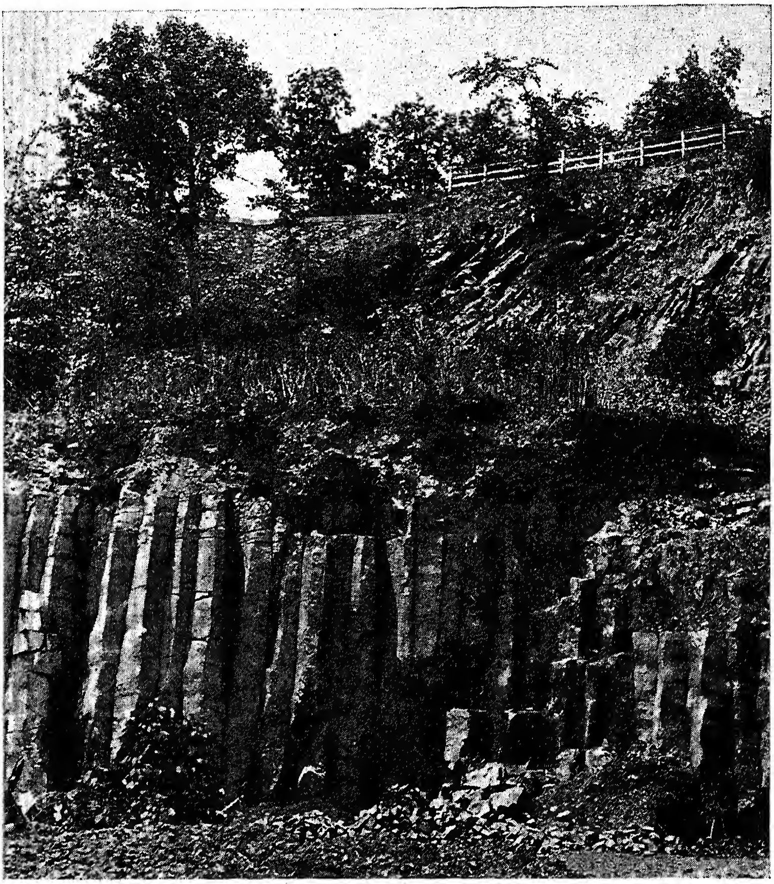

Fig. 321.—Columnar jointing in a Triassic intrusive, Watchung Mountain, Orange, New Jersey. (*Photograph by Iddings, U. S. Geol. Survey.*)

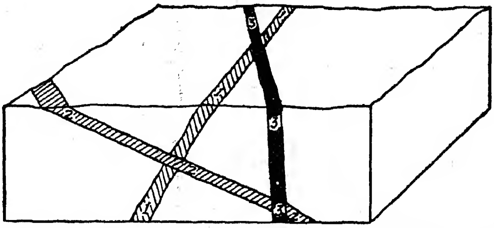

Fig. 322.—Diagram showing dikes of different ages. The oldest, 1, is cut by dike 2, and 1 and 2 are cut by dike 3.

shows a dike that is younger than rock 1 and older than rock 3. The determination of age relations of intrusives is an essential part of the systematic study of an area, and particularly it is important in areas containing mineral deposits, which commonly are related to the same magmatic reservoir that supplied the material of the intrusive. An intrusive that contains fragments of another rock is younger than the rock that supplied the fragments. When an intrusive alters another rock near its contact, it is younger than that rock. Certain intrusives are fine grained or glassy near their contacts with other rocks, whereas a few feet away from the contact they are more coarsely crystalline. The chilled margin is evidence that the intrusive rock is later than the cooling surface that chilled it. These relations are illustrated by Fig. 324. If one igneous rock is regionally metamorphosed and another one near by is not, the metamorphosed rock is the older; or if any regional change has taken place in one rock and the other rock is not affected, the altered rock is the older, for obviously both rocks would have been altered if both had been present when regional alteration took place.[1]

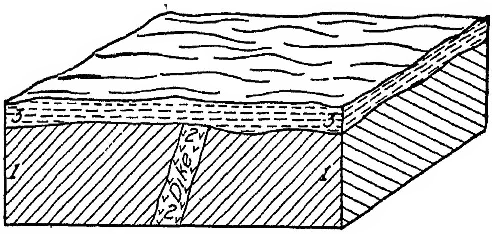

FIG. 323.—Diagram illustrating age relations of different rocks. A series of sediments, 1, after tilting are intruded by dike, 2. Subsequently these rocks are worn down by water. The area is submerged, and series 3 is deposited on the eroded surface. The area subsequently was raised above sea level.

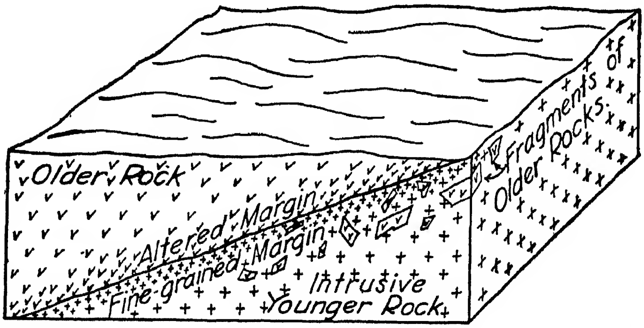

FIG. 324.—Diagram showing a younger igneous rock intruding an older igneous rock. The margin of the older rock is altered by solution from the younger rock, and fragments of the older rock are included in the younger rock. The margin of the younger rock is fine grained near the contact.

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[1] For references see page 321.

CHAPTER XII

IGNEOUS ROCKS

Classification of Igneous Rocks.—Igneous rocks have been classified in different ways. The principal rocks may be grouped on the basis of texture and of mineral content. They are, on the basis of texture, (1) fragmental rocks, (2) glassy rocks, (3) aphanitic or felsitic rocks, (4) porphyritic rocks, (5) granitoid rocks.

On the basis of composition, igneous rocks may be classified as acidic or basic. If silica is abundant in a rock, it is *acidic.* If the bases such as iron and magnesium oxides are abundant, the rock is *basic.* The acidic and basic constituents of a magma crystallize as characteristic minerals. Consequently the mineral composition may be used as a basis for classification (page 319).

A fragmental igneous rock is made up of fragments of igneous material. They are pyroclastic (fire-broken) rocks. Example, volcanic breccia.

A glassy rock is one that cools so rapidly that few or no crystals form. All or part is glass. Example, obsidian.

An aphanitic or felsitic rock is one whose mineral constituents are so small that they cannot be distinguished by the naked eye. Example, felsite.

A porphyritic rock is one with an aphanitic groundmass that encloses visible crystals. Example, quartz porphyry. A rock composed entirely of crystals some of which are considerably larger than others also has a porphyritic texture. Example, porphyritic granite.

A granitoid rock is one in which all the leading mineral constituents may be seen with the naked eye. These are called also phaneritic rocks. Example, granite.

Factors Causing Textural Variations.—The texture of an igneous rock reveals the manner of its formation. If cooling is slow, crystallization proceeds from few centers, and large crystals will form. If cooling is rapid, it proceeds from many centers, and small crystals form. If fluids are present, they tend to lower the temperature of crystallization; they promote diffusion or movement of material through the magma, and this permits the growth of larger crystals. A magma may be fluid at a temperature so low that it would solidify readily if the gases were not present. If such a magma flows out upon the surface, the gases escape, and the molten matter solidifies quickly to form an aphanitic rock or a glass. If some crystals had formed at depth before eruption, phenocrysts would be present, and the rock would be porphyritic. At considerable depths below the surface, fluids may be retained in the magma. If they are still present when crystallization begins, the mineral particles or molecules can move through the liquid magma more readily, and larger crystals or mineral grains are formed. In general, the basic magmas which are low in silica and high in iron and magnesia are much more liquid than the silicic magmas, and consequently the crystals or grains of minerals in general grow to larger size in basic lavas than in the more viscous siliceous magmas. This is not true in many deep-seated rocks, however, for pegmatites which have the largest crystals are of siliceous composition.

Some rocks possess a texture that suggests two periods of crystallization. In many of them closely spaced large crystals are found surrounded by a fine-grained or glassy groundmass. Where such textures are found in certain lava flows, it is believed that the large crystals were formed at considerable depths before the lava rose into the vent of the volcano. In many such flows the crystals are lined out by movement of the moving lava, which shows that the crystals had formed before the molten magma came to rest.

Textural variations are influenced also by the mode of occurrence of igneous rock masses. Since the texture of an igneous rock depends mainly on the rate at which a magma or lava solidifies, it follows that large intrusive masses will form coarser textured rocks than thin lava sheets or thin dikes and sills. In general, the rocks that occur as great batholiths are of granitoid texture, whereas dikes and sills are composed of fine-grained, or felsitic, rocks. Thin lava flows generally are glassy. Not all small bodies of rock, however, are fine grained. The volume of rock in a volcanic neck is not great, but notwithstanding the small volume, the rock commonly is coarse grained. The coarse texture is due to slow cooling. The constant upward passage of molten material heats the rocks surrounding the conduit of the volcano. When the magma in the conduit finally solidifies, it cools slowly because the surrounding rocks are hot.

Lava Rocks.—The term lava is used for the magma that flows out upon the surface and loses its volatile portions. The term is used also for this material after it has cooled and solidified (Fig. 306). Such rocks show certain structures not found in coarse-grained intrusive rocks.

Lavas commonly show banding; their materials are rarely uniform but include colored spots, gas cavities, etc. As they flow, they drag out the different kinds of material, and these form bands (Fig. 325). The lavas generally contain large amounts of gases when they are expelled from the volcanoes. As they rise to the surface, the pressure on them is relieved, and the gases expand. As a result the lava often becomes porous, or vesicular. On solidification it makes a froth-like rock, which is called *pumice,* or a rock with fewer and larger voids, which is called *scoria* (Fig. 24). Siliceous lavas, on account of high viscosity, form pumice more readily than basic lavas. Pumice is formed also before lavas are expelled from craters by the boiling and the frothing of the lavas in the craters. Vescicles that are filled by later deposition are called *amygdules.* By chilling effects, the base of a lava flow that rests upon a cooled surface may show more glass and a texture different from that of higher parts. The tops of flows generally are more porous than their bottoms. A few feet below the surface basic lavas commonly are free from vesicles. Either the pressure of the lava is sufficient to prevent the expansion of the gases contained in the lava, or all of the gases rise to the top of the flow. Lavas move while vesicles are forming, and these are drawn out so that layers of them are formed. This emphasizes the banding of the rock, and often the vesicles coalesce so that the openings are joined. Near the top of the flow the gases tend to rise, and thus the vesicles are elongated upward, forming pipes. These pipes are several

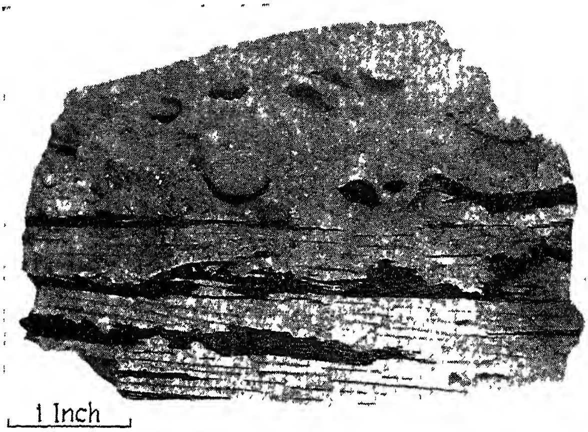

FIG. 325.—Photograph of a fragment of a rhyolite flow. The gray areas represent reddish brown rhyolite. The black bands are obsidian or black glass. The dark gray areas of the upper part of the figure show cavities formed by expanding gases.

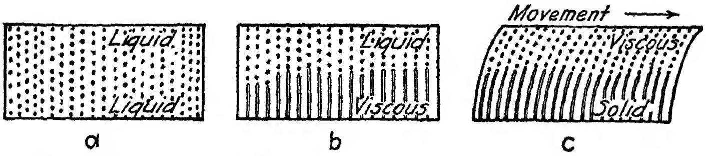

FIG. 326.—Drawings illustrating the development of bent-pipe vesicles. *a,* gas bubbles rising through a liquid lava; *b,* pipe-like vesicles forming in partly viscous lava; *c,* due to movement of the viscous lava, the pipe-like vesicles are bent in the direction of movement which is indicated by the arrow. Pipe-shaped amygdules are formed by subsequent filling of the vesicles.

inches long and have the shape of wheat straws. If the lava moves while they are forming, they are bent. The bottom of the flow drags on its base, and the top moves more than the bottom, so that the top of the pipe is bent over in the direction toward which the lava moves (Fig. 326). Even after the lava bed is steeply tilted, it is possible to ascertain the direction from which the lava came by the positions of the bends of the pipes. On account of their closely spaced cells, vesicular lavas and pumice are lighter than the solid rock. Pumice will float on

FIG. 327.—Pillow lava. (*Photograph by Francis Pettijohn.*)

FIG. 328.—Spherulites in rhyolite, Sawtooth Mountain, near Rhyolite, Nevada. (*After Ransome, Emmons, and Garrey, U. S. Geol. Survey.*)

water many days before it becomes waterlogged and sinks. *Scoriae* are cellular lavas coarser than pumice. They are also called volcanic cinders.

Certain flows that are broken by irregular curving joints are called *pillow lavas.* They are believed to have formed under water (Fig. 327).

Where gases are concentrated in glasses, they seem to promote crystallization, so that more small crystals are formed or the crystals are formed with different arrangements. Thus *spherulites* (Fig. 328), or small globular masses, are commonly developed in glassy rocks. In these, minute crystals of feldspar are arranged like rods radiating from a center.

Some lavas are very dense or aphanitic; that is, they do not contain crystals that are easily seen with the naked eye. Those that are dark colored are called *basalt.*

FIG. 329.—Volcanic bombs. (*Photograph by Grout.*)

Pyroclastic Rocks.—Pyroclastic rocks are those broken up by volcanic processes. They include the flow breccias and rocks that have been hurled from the craters of volcanoes. Volcanoes erupt intermittently, and lavas harden and fill their craters. For long periods some are quiescent. Lavas and gases accumulate below the solid seal of the crater; and when the seal is no longer strong enough to hold them in, the rock stopper gives way and a violent eruption follows. When molten lavas are thrown from craters, often they are drawn out into hair-like filaments of glass, known as *Pele's hair.* Small fragments are called *lapilli,* and larger and generally somewhat rounded ones are called *volcanic bombs* (Fig. 329). A large portion of materials thrown from craters consists of fragments of solid rock. These fragments range in size from fine dust to rock masses many feet in diameter. Coarser fragments are called *cinders,* and the fine dust-like particles are called *volcanic ash,* although they are not the residue of combustible material. The pyroclastic material finds lodgment on the surface and forms beds of tuff which include both ash and large fragments. In general, the tuff beds are very poorly stratified, although certain parts that have found lodgment in pools and lakes are layered as a result of assortment in the water. Great formations are built up by the material thrown from volcanoes. Near Ouray, Colorado, they cover scores of square miles, and in the Amphitheater, a deeply eroded canyon near Ouray, the series of tuffs are seen to be more than a mile thick. *Agglomerates* are deposits made up in part of large and small angular blocks of lavas.

In general, the volcanoes that yield siliceous lavas erupt more violently than those that yield basic lavas, and larger amounts of gases are associated with the siliceous lavas; consequently they generally yield the pyroclastic rocks in larger amounts, although this is not universally true.

As the top of a lava flow is more exposed, it will cool more rapidly than the lower part; consequently the top often will form a crust which rests on the lower part. Because of movement, the crust breaks into numerous blocks, and these blocks are largely light vesicular material. The masses of fragments floating on the lavas form great island-like bodies, large parts of which project above the liquid. At places essentially all of the surface of the flow is covered with the loose blocks. After cooling, the voids between the blocks form great systems of openings which extend for miles without interruption. Such openings later may serve as important channels for moving waters. The brecciated vesicular tops of lavas are well developed in basic flows. They are common features of the lava fields but are best known in the ancient basic lavas of Keweenaw Point,[1] Michigan. Light fragments of rock floating on the lavas are so numerous in certain lava streams that they hide from view completely the liquid flow; and when the stream advances on land, it moves forward like a "stone wall," only the solid matter being visible.

Lavas that contain numerous fragments of their broken crusts are called *flow breccias.* In the western United States there are series of lava flows extending over thousands of square miles which are made up of flow breccias in which the total mass of the fragments of the breccia is not much less than the lava matrix which cements them.

In Hawaii the relatively smooth lava surfaces are called *pahoehoe,* and the rough one is called *a-a.* These names have come into general use.

Porphyries.—A porphyry is an igneous rock containing visible crystals (phenocrysts) in an aphanitic, or fine-grained, groundmass. In general,

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[1] BUTLER, B. S., W. S. BURBANK, T. M. BRODERICK, L. C. GRATON, and others, The Copper Deposits of Michigan, U. S. Geol. Survey *Prof. Paper* 144, pp. 27–34, 1929.

the porphyries are intrusives, but most of them are intruded relatively near the surface or in marginal bodies or as smaller bodies that are cooled more quickly than the granitic rocks. They are therefore transitional between the granular rocks and the surface flows. Some lavas are porphyritic; they are made up of distinct crystals that are surrounded by a fine-grained or glassy groundmass (Figs. 330, 331). In certain lavas the larger crystals show parallel alignment due to movement of the lava during cooling (Fig. 331). This texture is common in basic lavas, particularly in diabases.

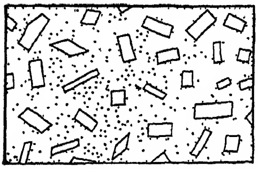

FIG. 330.—Sketch illustrating porphyritic texture. A fine-grained or glassy groundmass incloses visible crystals.

Although large crystals rarely form in siliceous lava flows, in certain thick basic flows moderately large crystals appear to have formed after the flow came to rest. Because the basic magmas are more fluid, larger crystals are formed in them than in the acidic rocks such as rhyolites. In certain basalts (trap rocks) large crystals have formed only in the interior parts of the flows. They are essentially absent from the tops and bottoms of the flows, because the latter cooled more quickly and the large crystals did not have opportunity to form. In certain flows in northern Michigan it has been found by Lane that the approximate position of a rock in certain flows may be estimated by the size of the grain developed by the larger crystals.

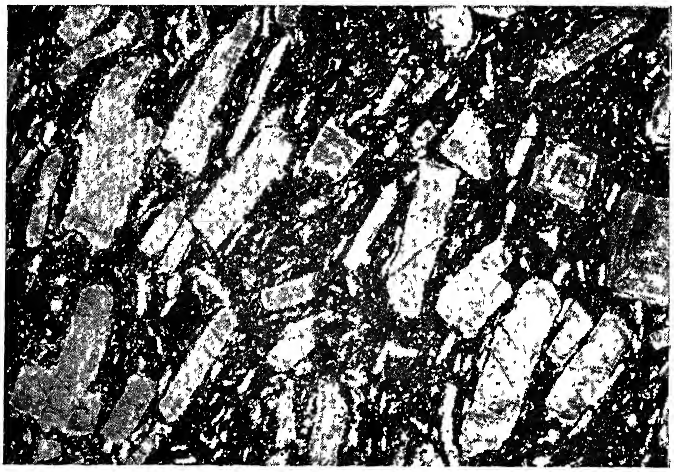

FIG. 331.—Photograph of a porphyritic syenite.

Granite and Rhyolite.—Granite, one of the most common of the igneous rocks, is made up of feldspar, quartz, and mica (Fig. 332). In some granites the mica is white mica (muscovite); in others it is dark mica (biotite). In certain granites, pyroxene or hornblende are present, and these may be more abundant than the micas. In a few granites the micas are absent altogether. Feldspar and quartz are generally light-colored; biotite, hornblende, and pyroxene are dark minerals. The color combination gives a salt-and-pepper effect like a white area spotted with black. A few granites are nearly all white. Some show a "graphic" intergrowth of minerals (Fig. 333). Granite is a deep-seated rock, and it is essentially free from glass.

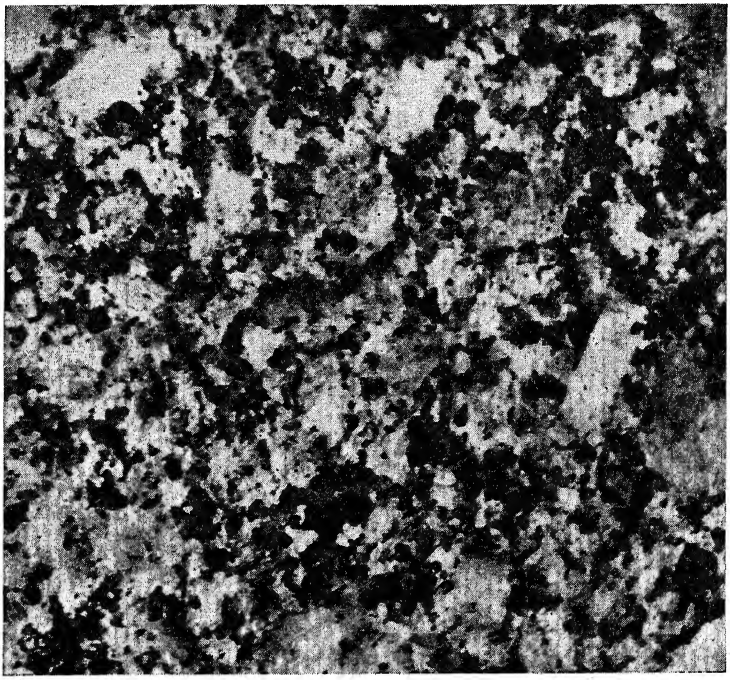

Fig. 332.—Granite.

A pegmatite is a rock generally composed of feldspar, quartz, and mica. In many pegmatites the crystals are much larger than the crystals of granite. Pegmatites are said to have been deposited by "aqueo-igneous" solutions—that is, by magmas unusually high in fluids.

If a magma like that which at depth solidified to form granite had flowed out upon the surface, the molten matter would have chilled before there was time for the molecules to organize themselves to form crystals. The resulting rock would be partly or entirely glassy; or if crystals were formed, generally they would be small. The rock would be *rhyolite;* or if it were essentially all glass, it would be called *obsidian.* As already stated, some rhyolites contain crystals that were formed at depths and were carried up with the molten rock. If a granite and a rhyolite are analyzed, they are found to have approximately the same chemical composition. Granite and rhyolite are the most siliceous (or acidic) of all the common rocks. Quartz is silica, and feldspars contain much silica. They are *salic* minerals. Micas, pyroxenes, and amphiboles are relatively low in silica. The dark minerals—biotite, amphiboles, and pyroxene—contain iron and magnesium and commonly are called ferromagnesian minerals or, briefly, *femic* minerals.

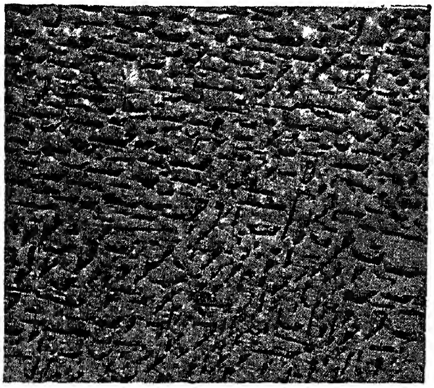

FIG. 333.—Graphic granite, Bedford, New York. Feldspar (light) and quartz (dark) are intergrown so that they resemble certain ancient characters used in writing. (*After Bastin.*)

Syenite and Trachyte.—If a less siliceous magma were to rise within the earth's crust and solidify at depth, the rock formed would be less siliceous than granite; if there were so little silica present that no quartz was formed but only feldspar and the dark minerals crystallized out, then the rock would be *syenite*, which has about the same mineral composition as granite but contains little or no quartz. If the same magma flowed out upon the earth's surface, it would form *trachyte*, which is the finely crystalline or glassy equivalent of syenite.

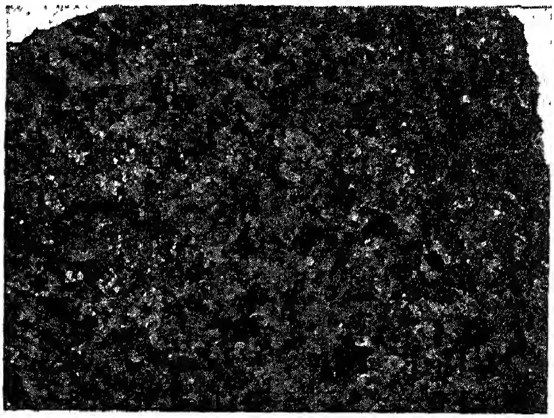

FIG. 334.—Gabbro.

Diorite and Andesite.—Diorite is a deep-seated crystalline rock made up of feldspar, which is mainly sodium and calcium feldspar, and one or more of the dark minerals—biotite, pyroxene, or amphibole. The flow equivalent of diorite is *andesite*. Normal diorite contains little or no quartz. If much quartz is present, the rock is *quartz-diorite*.

Gabbro and Basalt.—Gabbro, like diorite, is a deep-seated rock. It contains essentially the same minerals as diorite, but it has more ferromagnesian minerals and smaller amounts of feldspar (Fig. 334). Basalt is the surface-flow equivalent of gabbro. In some gabbros and in most basalts, olivine is present. Gabbros containing olivine are called *olivine gabbros.*

Peridotite.—Peridotite is composed chiefly of olivine.

COMMON IGNEOUS ROCKS

Mineral Variations	Textural Variations		
	Granular texture, rarely porphyritic (formed at great depths)	Porphyritic (formed at intermediate or shallow depths)	Aphanitic: glassy, felsitic, fine grained or porphyritic with fine-grained or glassy groundmass (formed at surface or shallow depths)
Acidic Rocks→ Salic minerals increase→	Granite: alkali feldspar and quartz with one or more of the minerals mica, amphibole, and pyroxene	Granite-porphyry	Rhyolite: some or all minerals of granite may be present. If all glassy, it is obsidian
	Syenite: alkali feldspar with one or more of the minerals mica, amphibole, or pyroxene	Syenite-porphyry	Trachyte: may be glassy in part; some or all of the minerals of syenite may be present
	Diorite: feldspar, mainly plagioclase with amphibole, pyroxene, or mica with two of them or all	Diorite-porphyry	Andesite: fine grained, may be glassy in part; some or all of the minerals of diorite may be present
←Basic Rocks ←Femic minerals increase	Gabbro: feldspar, mainly plagioclase, with pyroxene or amphibole or both; commonly mica. Generally contains more dark minerals than diorite. Some olivine commonly present	Gabbro-porphyry	Basalt: same minerals as in gabbro but finer grained. Olivine generally present; if glassy, it is basaltic glass; if scoriaceous, scoria
	Peridotite: olivine, generally with pyroxene or commonly with amphibole. Some contain a little feldspar	Peridotite-porphyry	Basalt: very rare. Same minerals as peridotite

Magmatic Differentiation.—Magmatic differentiation is the process in the operation of which magmas of supposed uniform composition have separated and on cooling have formed rocks of different compositions. At many places deep-seated igneous rocks are found grading one into the other (Fig. 335). Thus a light-colored granite will grade into a darker diorite or into a gabbro. Neither rock intrudes the other, and therefore they are believed to be of the same age. When magmas rise into the earth's crust, it is improbable that one part of the magma is very different from another. If they were different originally, they would tend to become uniform by mixing during their ascent. It is believed that both rocks, as illustrated by Fig. 335, were derived from the same magma and that the magma separated to form into two different rocks on cooling. In certain laccoliths one may observe that the rock in the lower part of the laccolith is different from that in the upper part. The heavier rock generally is below the lighter one, and this suggests that the differentiation is due to gravity. The dark, heavy crystals such as magnetite and pyroxene are more abundant in the lower rocks, whereas quartz and feldspar are more abundant in the upper ones. The inference is warranted that the heavier crystals or heavier molten material have sunk to the bottom of the chamber and that the lighter ones have risen to the top.

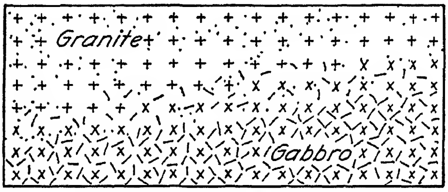

Fig. 335.—Diagram showing a basic rock, gabbro, grading into granite. The two rocks are of the same age and are differentiates of the same magma.

In the great batholiths, differentiation probably is carried out on a vastly larger scale than in the laccoliths with magmatic chambers of restricted size. The batholiths, however, are so large that the operation of the processes is more difficult to interpret, and moreover the floors of batholiths do not come under observation.

Rise of Magmas.—The behavior of earthquake waves shows that the earth is essentially solid. Much of it probably would be molten if it were not for the great pressure due to the weight of the rocks themselves. When the pressure is relieved by folding or faulting, rocks melt to form magmas which rise by melting, thrusting, or stoping their way up. This theory is supported by the fact that great igneous activity generally is coincident with faulting and folding. That is true at many places throughout the earth and is well shown in the western part of the United States.[1]

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[1] Butler, B. S., C. D. Hulin, A. C. Waters, C. P. Ross, W. S. Burbank, T. S. Lovering, and H. Schmitt, Lindgren Volume, pp. 198–326, 1933.

CHAPTER XIII

DIASTROPHISM

Diastrophism includes all movements of parts of the solid earth with reference to each other. These include the rapid vibrations such as those that result from earthquakes and the slower movements continued for longer periods that result in the formation of mountain ranges and in the lifting up or sinking down of areas of continental dimensions. Evidences of diastrophism abound. Major earthquakes are recorded at an average of about one per week, and accurate readings of levels established on certain coasts show that at places the sea lately has encroached upon the land so that deep river channels formed on the continents are now beneath the waters of the sea.

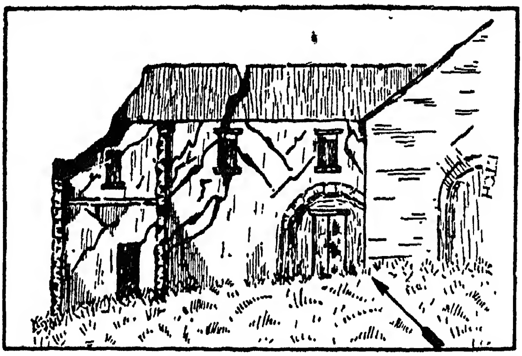

FIG. 336.—Sketch showing a house rent by earthquake shocks. The arrow shows the direction from which the earthquake wave emerged. (*Redrawn from Mallet.*)

EARTHQUAKES

Earthquakes are short and rapid vibrations that result from the deformation of the earth's outer shell mainly by folding or faulting or by the adjustment of rocks along faults. Earthquakes widely recorded occur on an average of one in 14.5 hours. They are among the most destructive of natural phenomena and often are even more terrifying than the eruptions of volcanoes, since the earthquake shock takes place on the ground, which generally, from childhood, men have regarded as stable. The earth's crust is almost continuously in a state of tremor. Some tremors are minor vibrations which result from traffic, from the blast of mines, or from the fall of rock.

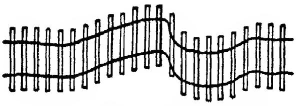

FIG. 337.—Railway deformed by Charleston earthquake of 1886. (*After Dutton, U. S. Geol. Survey.*)

Intensity of Shocks.—Earthquakes affect large areas, and many of them are recorded over the entire earth. There is generally an area, however, within which the shocks are most severe. Such a region usually is a broad belt, and as it is assumed to lie above a fault or fold where the shock focus, or centrum, is located, it is called the *epicentrum*" (above the center). In such areas houses are destroyed (Fig. 336), trees are split, railway tracks and piers are bent (Figs. 337, 338), water pipes and cables are broken, roads and fences are offset. The earth opens, forming great cracks and fissures. The sides of hills and bluffs slip away as landslides (Fig. 339). Where stiff clay covers wet sand, fractures open in the clay, and fountains of water and sand spurt out. Craterlets form openings in the clays and sand hills, and sand "blows" result where the sand is thrown out over the surface. In loose, wet ground or "made" ground particularly the shocks are severe (Fig. 340). The intensity of shocks away from the epicentrum dimin-

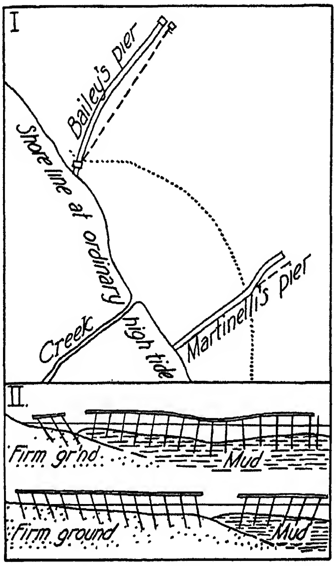

FIG. 338.

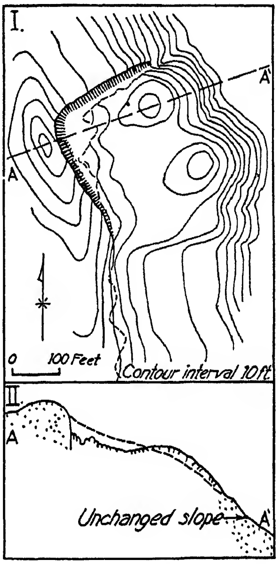

FIG. 339.

FIG. 338.—*I*, plan showing the piers at Inverness, Tamales Bay, about 24 miles northwest of Golden Gate, San Francisco, California, near San Andreas fault. The broken lines show positions of piers before the earthquake of 1906. The solid lines show locations of piers after the earthquake. The dotted line indicates the limit of the area of the shifting of the bottom of the bay during the earthquake. *II*, diagrams showing deformation of piers in I. Bailey's Pier is above and Martinelli's Pier below. (Vertical scale exaggerated.) The piers were broken by movement of the sand of the bay. Horizontal lines show water level. (*After California Earthquake Commission.*)

FIG. 339.—*I*, map of landslide area near San Francisco Bay, California, about 4 miles east of San Pablo showing the area of slide during California earthquake of 1906. Scale and contour intervals are approximate. *II*, cross section of landslide shown in *I* along the line *AA*. The land slumped below a fault indicated on the cross section. The broken line indicates the slope before the landslide. (*Data from F. E. Matthes and E. S. Larsen.*)

ishes rapidly. A few score miles away the destructive effects of the shocks are slight. The belts which show similar intensities as recorded by the effects of the shocks are *isoseismals* (same shocks). Scales

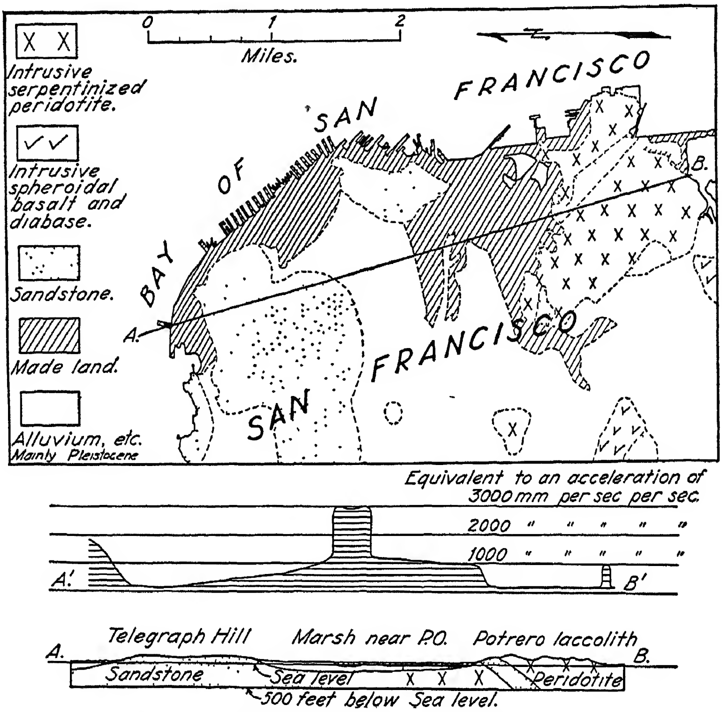

FIG. 340.—Geologic map of part of San Francisco and (below) a cross section showing by elevation of shaded part the relation of places of greatest intensity of the shock to the areas of "made land" and alluvium in the earthquake of April 18, 1906. The shock was least intense in solid rocks—the sandstone and peridotite shown in the lower section. (*Redrawn from data presented by California Earthquake Commission.*)

for classifying areas according to the intensity of shock have been proposed. The one generally used is the Rossi-Forel scale, shown abridged below.

ROSSI-FOREL SCALE OF INTENSITIES OF EARTHQUAKE SHOCKS[1]

1. *Microseismic Shock.*—Recorded by a single seismograph or by seismographs of the same model but not by several seismographs of different kinds; the shock is felt by an experienced observer.
2. *Extremely Feeble Shock.*—Recorded by several seismographs of different kinds; felt by a small number of persons at rest.
3. *Very Feeble Shock.*—Felt by several persons at rest; strong enough for the direction or duration to be recorded.
4. *Feeble Shock.*—Felt by persons in motion; disturbance of movable objects, doors, windows, cracking of ceilings.

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[1] HECK, N., and R. R. BODLE, United States Earthquakes, 1928, U. S. Dept. Comm. *Ser.* 483, p. 2, 1930.

5. *Shock of Moderate Intensity.*—Felt generally by everyone; disturbance of furniture, beds, etc.; ringing of some bells.

6. *Fairly Strong Shock.*—General awakening of those asleep; general ringing of bells, oscillation of chandeliers, stopping of clocks, visible agitation of trees and shrubs; some startled persons leave their dwellings.

7. *Strong Shock.*—Overthrow of movable objects, fall of plaster, ringing of church bells; general panic without damage to buildings.

8. *Very Strong Shock.*—Fall of chimneys, cracks in the walls of buildings.

9. *Extremely Strong Shock.*—Partial or total destruction of some buildings.

10. *Shock of Extreme Intensity.*—Great disaster; ruins; disturbance of the strata, fissures in the ground; rocks fall from mountains.

Seismographs.—A seismograph (Fig. 341) is an instrument used for recording vibrations of the earth's crust. A heavy weight suspended by a wire will remain essentially at rest for a time owing to the inertia of the weight. It moves very slowly, whereas the earth below it moves rapidly. A stylus is attached to the suspended weight, and either a ribbon of paper or a cardboard disk is moved by clockwork against the stylus (Figs. 342, 343). Time intervals are laid off on the paper, and the record on the strip shows the amplitude of movement and the time of movement. Such a record is a seismogram. If the earth is at rest, the record traced on the paper strip is a straight line. If the earth is in a state of mild tremor, the recording line shows minute undulations. When an earthquake shock reaches the place where the seismograph is set up, the increased vibration of the earth is recorded on the paper. This movement is greatly magnified by a lever, and the pendulum and arms are arranged so that one of them records vertical movements, another records movements in a north and south direction, and a third records those in an east and west direction. The three records show the movements in three directions. Seismograms are shown in Figs. 342 and 343. By using such records a model may be built up showing the path of a particle of the earth where the record was taken. This was done by Sekiya with records of an earthquake in Japan, January 15, 1887. Figure 344 shows the path of a particle during 20 seconds of the earthquake. The model shows the directions of movement only; the length of the path is considerably enlarged. Certain seismographs are constructed so that a beam of light is thrown on a sensitive film and the movements of the earth are recorded by photographic processes.

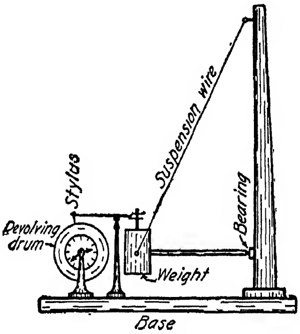

FIG. 341.—Horizontal-pendulum seismograph. The base of the instrument is mounted on solid rock. The suspended mass of the pendulum remains essentially stationary when the ground oscillates. The vibrations are magnified by means of a lever and are recorded on the paper strip of the revolving drum which is driven by clockwork.

The seismogram shows the time of vibration and the period of vibration, but the amplitude generally is greatly enlarged. It does not show the actual amount of movement of the earth. The time of vibration near the source of the earthquake is generally short. In the San Francisco, California, earthquake of 1906, most of the seismographs in the epicentral area were so violently shaken that their records were imperfect or useless. A section of a record on the Ewing seismograph at Lick observatory, Mount Hamilton, about 55 miles from San Francisco and about 25 miles from the source of the California earthquake is shown by Fig. 342. This shows that the oscillations of the earth occurred at intervals of about 2 seconds. The shock was recorded at Irkutsk, Siberia, 5,100 miles away, where the earthquake wave had passed through the earth. There the vibrations were a minute or more apart. In general, the movements are slower the farther the seismograph is from the source, and the destructive effects are much less.

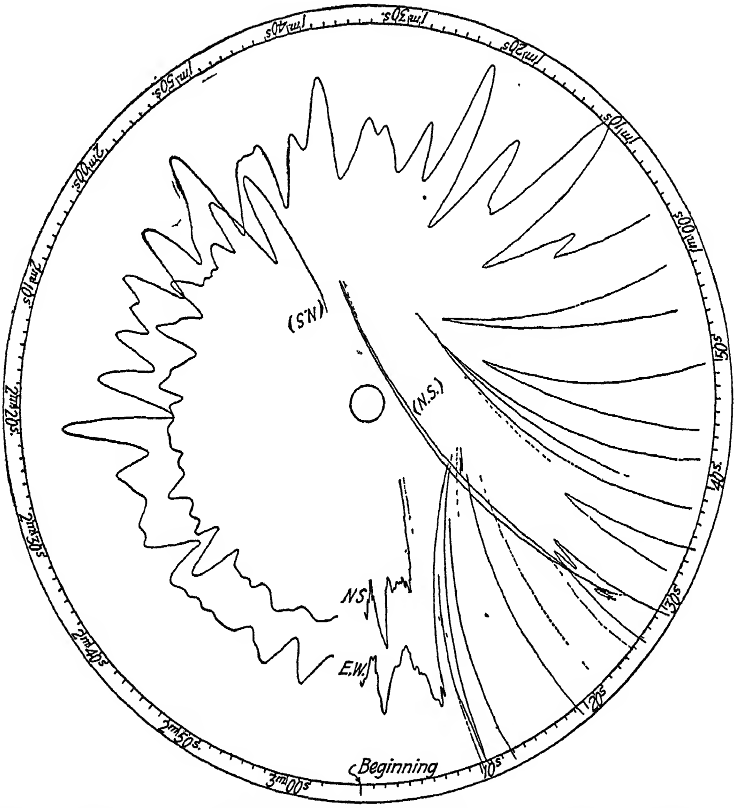

FIG. 342.—Part of a seismogram of the San Francisco earthquake of April 18, 1906, registered at Mount Hamilton, California, 55 miles southeast of San Francisco and 25 miles from San Andreas fault. The record was made on a Ewing three-component seismograph. The vertical element is omitted. Compare the speed of vibrations with those shown by Figure 343. (*After Report of California Earthquake Commission*, 1906.)

By utilizing the directions of movements of bodies and the positions of cracks (Fig. 336) it is possible to estimate the direction of the earthquake wave at the place of emergence and to estimate the depth at which the shock originated. This method was first used by Mallet. It is subject to certain limitations, but it has supplied valuable information. By using this and other methods calculations show that many earthquakes originate at depths between 5 and 25 miles below the earth's surface.

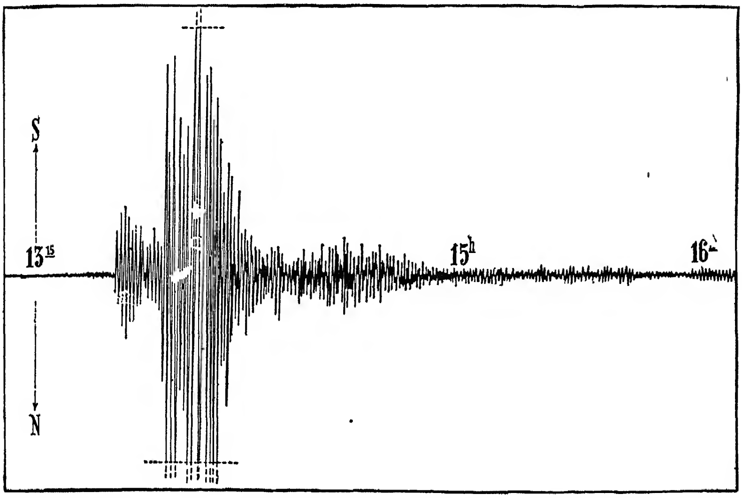

FIG. 343.—Seismogram of the San Francisco earthquake of 1906 registered by a horizontal pendulum seismograph at Irkutsk, Siberia. The oscillations recorded are made by earthquake waves that have passed deep into the earth. Note that the oscillations are much slower than those recorded in Fig. 342 near the source of the earthquake. (*After Report of California Earthquake Commission*, 1906.)

Location of the Epicentrum.—The compressional wave near the earth's surface travels about 7.5 kilometers per second, and the transverse wave travels about 4.2 kilometers per second (Fig. 345). Both waves travel faster with depth on account of the increase of elasticity with depth, but the fast wave is always about 80 per cent faster than the slow wave. The seismograms show the times of arrival of both slow and fast waves (Fig. 346). The difference in time of arrival of the two waves is 80 per cent, or four-fifths, of the time the slow wave has traveled. If the difference of time of arrival of two waves at St. Louis is 274 seconds (Fig. 346) and the slow wave travels 4.2 kilometers per second, there

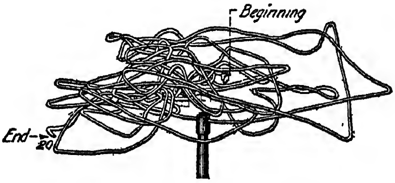

FIG. 344.—Diagram showing the path of a particle during an earthquake in Japan, January 15, 1887. The model by a bent wire shows position of a particle at 0, which is one end of the wire, and 20 seconds later at 20, marked "end" on figure. This model was made by Sekiya from seismograms. It is noteworthy that the particle moved in all directions at a very rapid rate.

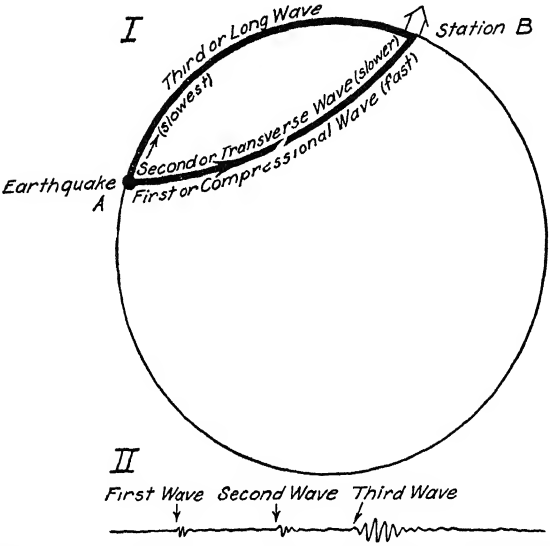

FIG. 345.—Diagram showing difference in the times of arrival of various types of seismic waves at seismograph station. (*After N. H. Heck.*)

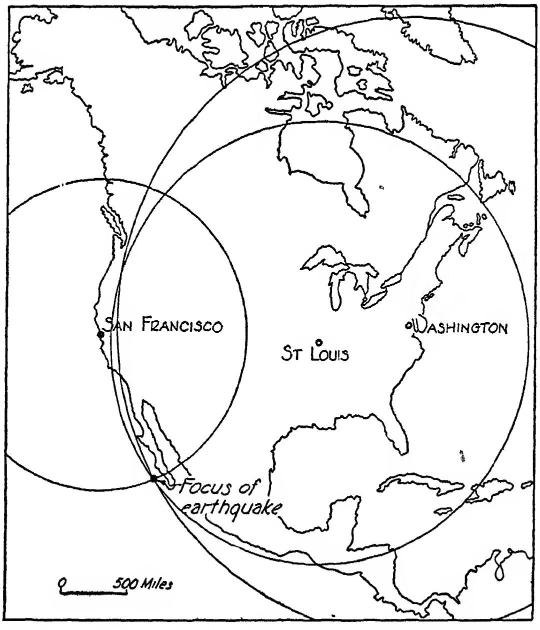

FIG. 346.—Map showing the method of locating the focus of an earthquake from seismograph records at three stations. The distances from San Francisco, St. Louis, and Washington are determined from the time elapsed between the arrival of the first and second seismic waves. Using the calculated distances as radii of circles drawn about the stations, they intersect near the coast of southern California.

would remain 1,150.8 kilometers to be traveled by the slow wave when the fast wave has reached St. Louis; 1,150.8 kilometers is about four-ninths of the total distance, 2,589 kilometers, which is the distance of the epicentrum of the earthquake from St. Louis.

If a circle is drawn with St. Louis as a center and a radius of 2,589 kilometers, the place of the earthquake will be on that circle. If seismograms at Berkeley and Washington are used and circles similarly are drawn from them, the place where the three circles intersect will be the place of the earthquake. In the ideal earthquake figured on page 328, the place of the earthquake is off the coast of Lower California (Fig. 346).

The study of seismograms shows that about 90 per cent of earthquakes originate near the earth's surface. But about 10 per cent of them originate at much greater depths, and a few at depths of 600 kilometers, or about 375 miles. Even at that depth below the earth, rocks slip and shear.

Nature of Movements.—As shown by seismographs, the earth moves in all directions. The movement during an earthquake shock, however, is small. The actual movement of the surface of the ground in a horizontal direction is often less than 0.1 inch, and the vertical movement is still less. When the horizontal movement reaches 0.5 inch, it becomes dangerous; and when it reaches 1 inch, it is generally disastrous. Movements of several inches have been recorded, but such movements seem to be rare except in loose material. Loose objects at the surface of the earth may move great distances. The nature of this movement may be illustrated by striking a solid table near a marble. The marble is thrown many inches, although the table moves only a fraction of an inch.

It is in sand, alluvium, and "made ground" that movement is greatest and earthquake shocks are most disastrous. In such areas the movement reaches a foot or more, and the ground heaves and swells like a disturbed bowl of jelly. Branches of trees interlock, men are thrown to the ground. In the New Madrid, Missouri, earthquake of 1811 it is said that earth waves several feet high were propagated across the country.

Destructive Effects and Their Prevention.—In many earthquake regions nearly every house in the areas of the epicentra is damaged and many are destroyed. The destructive effects result from the falling of roofs and chimneys and from the breaking of foundations and walls. A so-called earthquake-proof house, according to Milne, should be constructed with rafters running from ridge pole to floor sills and with iron straps and sockets replacing mortises and tenons. Light roofs and chimneys are recommended. Examinations of ruins of the San Francisco earthquake of 1906 showed that many of the bricks had been laid dry in mortar. They were poorly bonded, had clean surfaces, and the mortar showed little adhesion. Walls laid in cement with wet brick stood the test of the earthquake better. Reinforced concrete houses have proved relatively stable. The most secure house is one that will

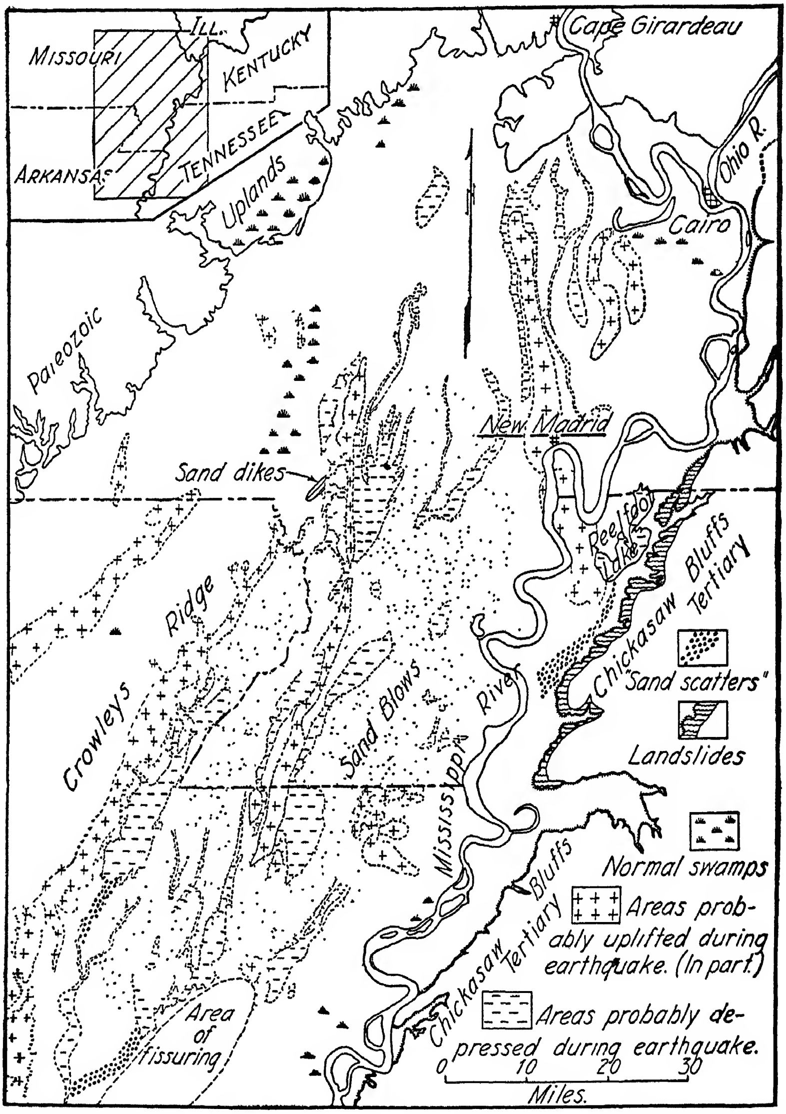

FIG. 347.—Map of part of region affected by the earthquakes of 1811 near New Madrid, Missouri. (*After Fuller, U. S. Geol. Survey.*)

move as a unit. For large structures soft ground should be avoided. River bluffs and sites near deep excavations are undesirable.

New Madrid Earthquakes.[1]—In the region near New Madrid, southeastern Missouri (Figs. 347, 348), a great series of earthquakes

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[1] FULLER, M. L., The New Madrid Earthquake, U. S. Geol. Survey *Bull.* 404, pp. 1–119, 1912.

began December 16, 1811, lasting more than a year. The region is part of the flood plain of the Mississippi River and is underlain by deep alluvium. The ground rose and fell in earth waves like the low swell of the sea, breaking timber, destroying houses, and moving trees. Landslides swept down from bluffs. Considerable areas were uplifted, and others were sunk. Numerous fissures, some of them 700 feet long and 20 feet wide, were formed. The shocks were felt over wide areas in the Mississippi and Ohio valleys. This earthquake probably was due to movement accompanying the settling of alluvial beds recently deposited on the flood plain of the Mississippi River, and this movement may have been attended by faulting below the flood plain (Fig. 348).

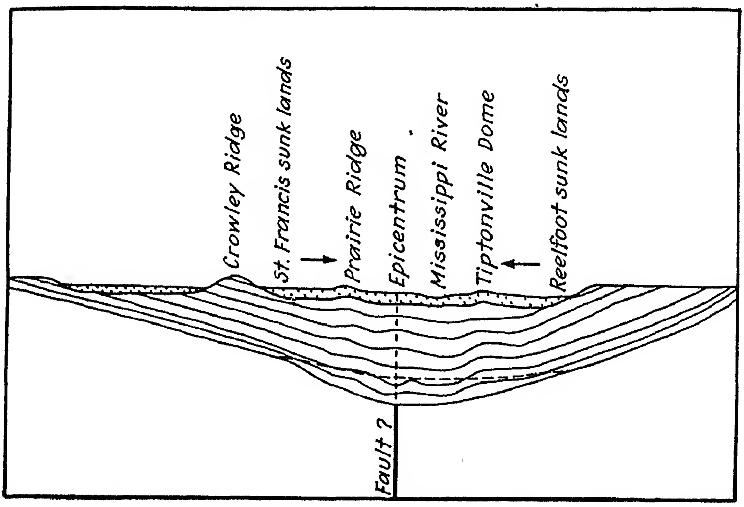

FIG. 348.—Diagrammatic cross section showing possible cause of doming of alluvial beds of Mississippi River flood plain during earthquake near New Madrid, Missouri, in 1811.

Charleston Earthquake.[1]—One of the most destructive earthquakes in the United States was that at Charleston, South Carolina, August 31, 1886 (Fig. 349). Nearly all the buildings of the town were damaged or destroyed, and 27 lives were lost. Railway tracks were bent (Fig. 337), and a train was overthrown. The belt most violently shaken was about 12 miles west of Charleston in an area of forest. During the earthquake, fissures opened and craterlets formed near this area at places shown by Fig. 349. Fissures discharged water in jets as high as 20 feet.

The earthquake was felt over most of the eastern part of the United States, the intensity of the shock decreasing with distance from the epicentrum, as is indicated by Fig. 350. The focus, or place of origin, of the earthquake was estimated by Dutton to be from 8 to 12 miles deep.

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[1] DUTTON, C. E., The Charleston Earthquake of August 31, 1886, U. S. Geol. Survey, 9*th Ann. Rept.*, pp. 203–528, 1889.

California Earthquake.[1]—On April 18, 1906, California (Fig. 351) was shaken by the most severe earthquake in the history of the state. People were killed, buildings destroyed, and in a fire that followed the earthquake a large part of San Francisco was burned. Piers in harbors near by were destroyed (Fig. 338), and landslides occurred (Fig. 339). The first snap or movement of the earth was recorded at the observatory of the University of California and was found to be 3 inches in a horizontal and 1 inch in a vertical direction. Quickly following the first snap were

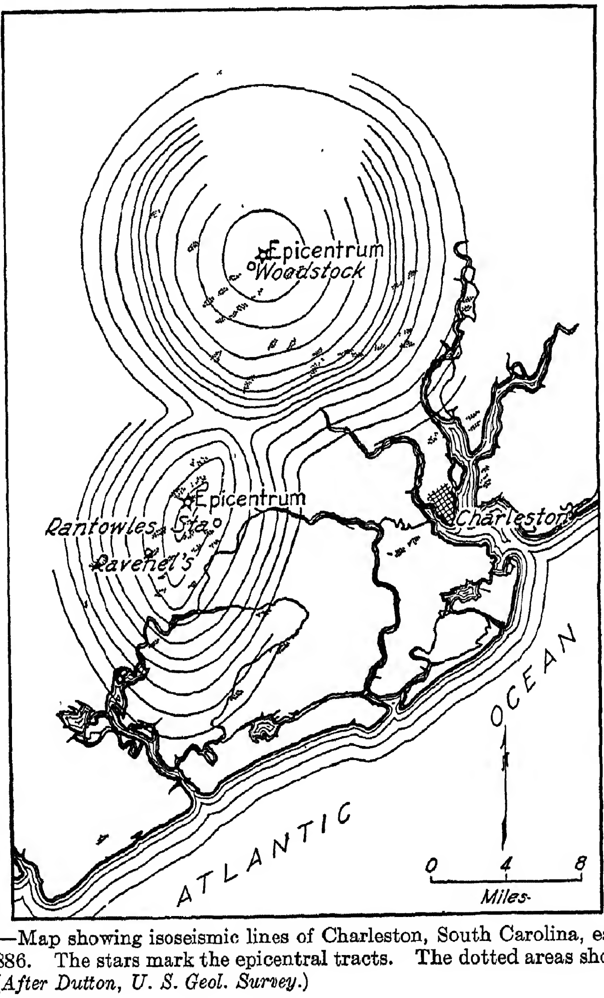

FIG. 349.—Map showing isoseismic lines of Charleston, South Carolina, earthquake of August 31, 1886. The stars mark the epicentral tracts. The dotted areas show regions of craterlets. (*After Dutton, U. S. Geol. Survey.*)

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[1] GILBERT, C. G., R. HUMPHREY, J. S. SEWELL, and F. SOULE, The San Francisco Earthquake and Fire of April 18, 1906, U. S. Geol. Survey *Bull.* 324, pp. 1–169.

LAWSON, A. C., and others, *Rept.* Calif. Earthquake Investigation Commission.

rebounds over the greatly disturbed area on both sides of a fault line. These movements, or "temblores," brought down chimneys and towers. The shocks were felt from Coos Bay, Oregon, to Los Angeles and were recorded at Washington, D. C.; at Potsdam, Germany; Irkutsk, Siberia; and other places.

The San Andreas fault extends S.35° E. from Point Arena 400 miles or more (Fig. 351). Study of this fault zone showed that movements had taken place about 270 miles along the fault. The fault has been a place of movement for long ages. Preceding the earthquake, movement on the fault probably resulted in bending the rocks on the sides of the fault. When the strain reached a degree beyond which it could no longer be borne without rupture, there was an abrupt movement along the ancient fracture. The relieving snap was like a blow upon the rocks along the fault.

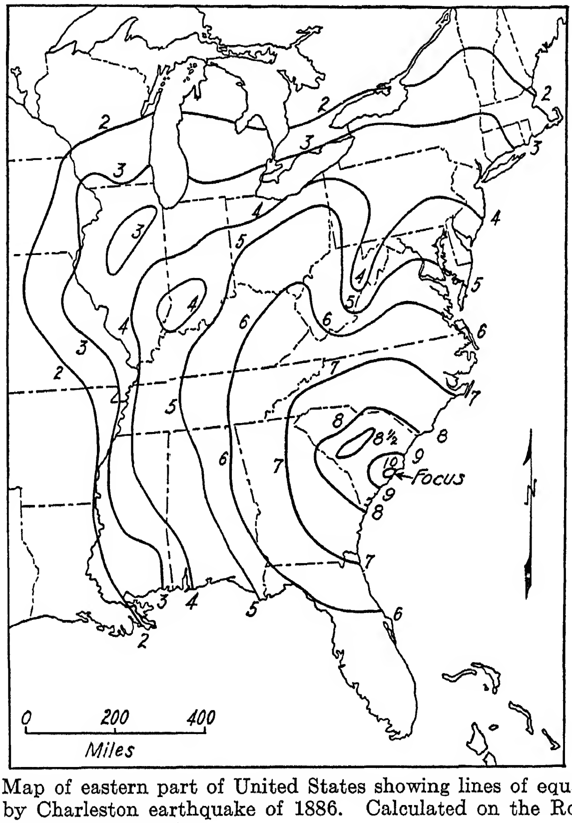

FIG. 350.—Map of eastern part of United States showing lines of equal intensities of shocks caused by Charleston earthquake of 1886. Calculated on the Rossi-Forel scale. (*After Dutton, U. S. Geol. Survey.*)

The fault that caused the California earthquake is approximately vertical, and the movement along it was approximately horizontal.

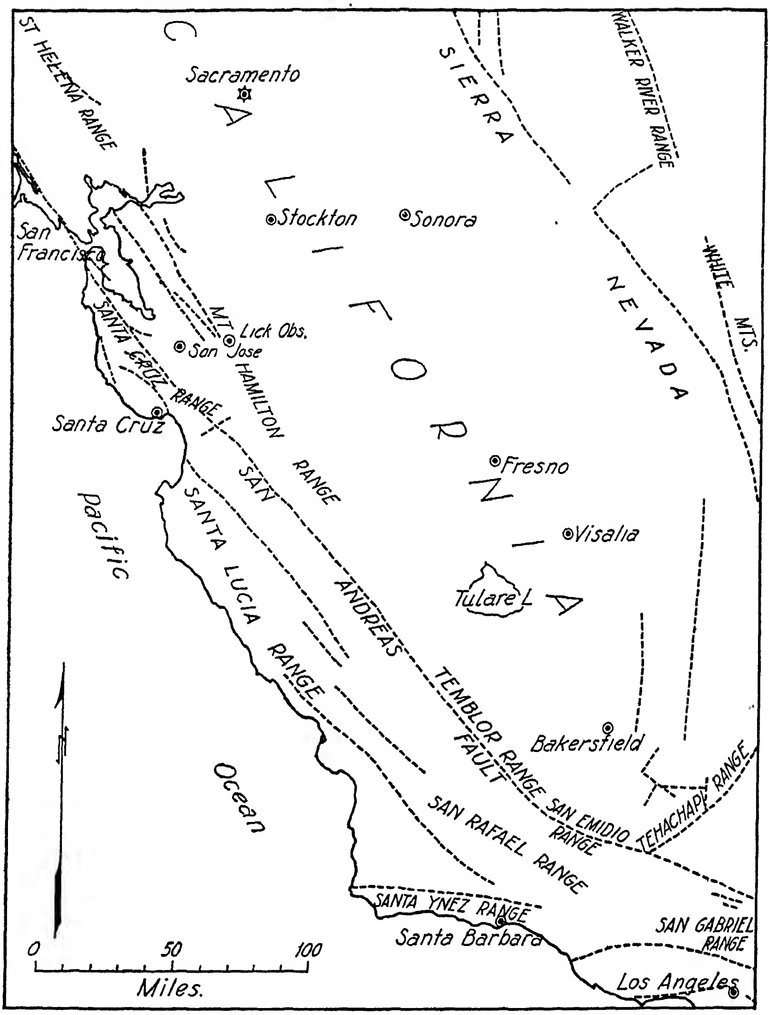

FIG. 351.—Map of part of California showing major faults. The earthquake of April 18, 1906, was caused by movements along the San Andreas fault, which passes through San Francisco.

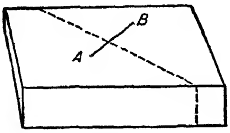

FIG. 352.

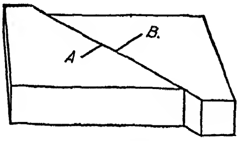

FIG. 353.

FIG. 352.—Diagram showing the direction and amount of movement on the San Andreas fault, which caused the California earthquake of 1906. The broken line, left figure, represents the fault. *AB* is a straight line crossed by the fault.

FIG. 353.—The same as Fig. 352 after faulting. The horizontal movement was from 2 to 20 feet, and the vertical displacement from 0 to 3 feet. The southwest block was elevated. (*After G. K. Gilbert, U. S. Geol. Survey.*)

This movement is shown by Figs. 352, 353. In general, the effects were most severe nearest the fault, although they varied considerably with the character of the rock, being severest on made ground (Fig. 340). Where foundations were laid on piles driven into the ground, the destructive effects were generally slight, even in the area of made ground. The buildings that were most affected by the shocks were on raft-like foundations on the loose mantle rock or fill.

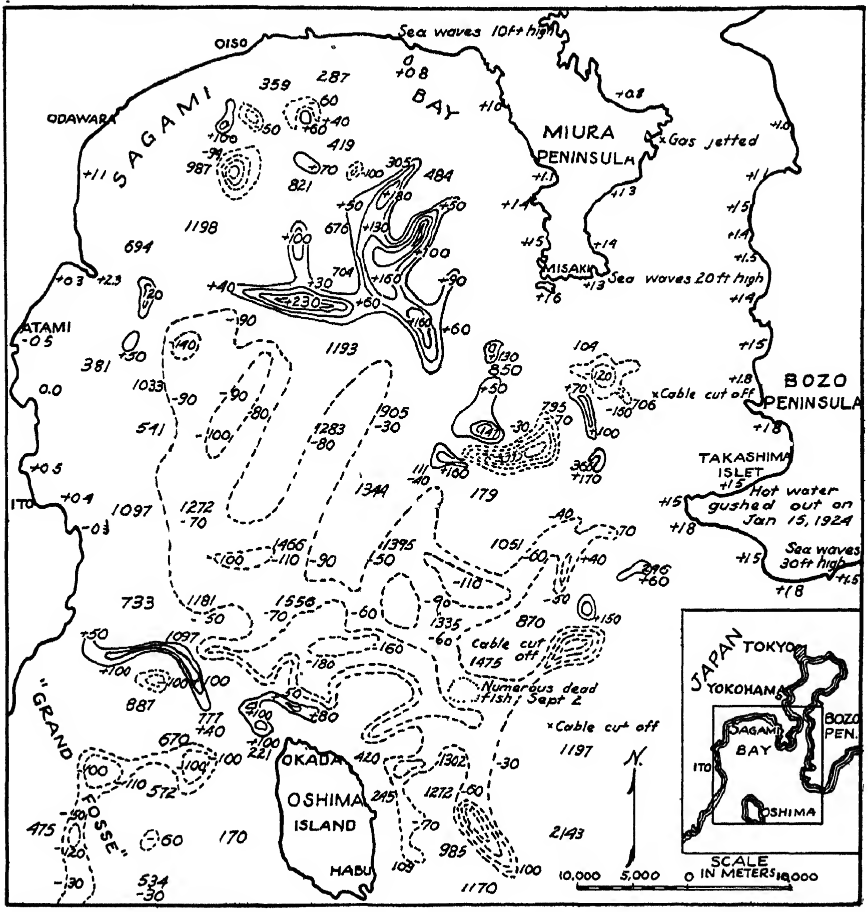

FIG. 354.—Map of Sagami Bay, south of Yokohama, Japan, showing the deformation of the floor of the bay during the earthquake of September 1, 1923. Numbers without prefixes show depths in meters before earthquake. Plus signs show elevations, minus signs show depressions during earthquake. Solid contour lines show elevation; broken contour lines show subsidence. (*Based on map issued by Bureau of Social Affairs, Japan.*)

Sagami Bay.—On September 1, 1923, a great earthquake destroyed Yokohama and a large part of Tokio, Japan. At 11:58:44 a.m. a violent shock shook the houses and other structures. Other shocks followed, and fire was kindled at scores of places. Owing to the bursting of water mains the fire could not be extinguished. At 4 p.m. a high wind swept the city. Sea waves destroyed shipping along the shores. The bottom of Sagami Bay, which is about a mile deep, had been mapped by soundings before the earthquake, and it was mapped also after the earthquake. As shown by Fig. 354, the movements of the sea bottom at places amounted to hundreds of feet.

Long Beach, California.—At Long Beach, California, there was a shock March 10, 1933. At short distances away it was not a strong shock as recorded by seismographs, but it was severe locally, and since there is deep alluvium over much of the area, there was heavy damage and some loss of life. The epicentrum was a little offshore. There was no sea wave and no notable disturbance of water, although vessels near by felt the tremors.[1]

Helena, Montana.—Some earthquakes consist of a large number of small shocks. These are *swarm earthquakes.* At Helena, Montana, from October 18, 1935, to January 1, 1936, there were nearly 1,300 shocks; a strong shock on October 18, one on November 28, and one on December 31 were destructive. Forty houses were ruined, 200 considerably damaged, and 80 per cent of the homes in the city suffered some damage. In mines near by, the shock was felt strongly, but no damage was done to them.[2]

Submarine Earthquakes.—The sea bottom is subject to earthquakes, as is the land. Water is almost incompressible, and sudden movements affect it much as they would a rigid body. Fish are killed during earthquakes, as they would be by a heavy explosion. Ships can better withstand the shocks, because they are free to move upward, yet shocks are felt on ships. Sailors have been thrown upon the decks, and submarine cables have been broken at times of earthquakes.

Sea Waves.—During certain earthquakes great waves are generated in the regions of violence where these are near coast lines. Formerly these were incorrectly designated as tidal waves. Often they are more destructive than the earthquakes. On August 13, 1868, the west coast of South America was shaken from Quayaquil, Ecuador, to Valdivia, in Chile, the most violently shaken region being in the neighborhood of Arica,[3] a city on the west coast, where many buildings were destroyed. A few minutes after the destructive shock the sea slowly receded from the shore, and ships anchored in 42 feet of water were left dry. Later the water returned as a great wall, caught up the ships, and swept them inland as if they had been chips of wood. The United States steamer *Wateree* was carried inland ¼ mile with little damage and left ashore.

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[1] Heck, N. H., Earthquakes, p. 143, 1936.

[2] Heck, N. H., Earthquakes, pp. 46, 141, 175, 1936.

[3] Dutton, C. E., Earthquakes, p. 281, 1904.

On November 1, 1755, a great earthquake and sea waves destroyed Lisbon, Portugal. The waves were 30 to 60 feet higher than the highest tide and came about one hour after the town had been shattered by the earthquake. At first the sea withdrew, the withdrawal of the water being followed soon after by the greatest wave. The water waves reached the coast of America in 9½ hours. Earthquake and great sea waves attended the eruption of the volcano Krakatao in August, 1883. The waves were 100 feet high and, moving inland from the coasts of Java and Sumatra, destroyed towns. In 1923 great sea waves accompanied the disastrous earthquake of Sagami Bay, Japan. The waves were noted 4,000 miles from their source.

Many earthquakes are not attended by sea waves. Kluge states that of 15,000 earthquakes observed on coast areas only 124 were accompanied by sea waves. During certain earthquakes great warpings of the sea bottom probably take place. That was true of the great earthquake in Japan in 1923 (Fig. 354). A sudden movement of this character would set in motion a great body of water. The water rushing to a depression of the sea would build up a great wave, and on a shallow, sloping beach it would pile up on a low coast. The most destructive sea waves are those which approach gradually shallowing water. As was pointed out by Darwin, Valparaiso, Chile, which is on the edge of deep water, has never been overwhelmed, although it is in a region where great destruction has resulted from sea waves.

Distribution of Earthquake Areas.[1]—Although earthquakes are recorded over the entire earth, their places of origin are in limited areas. The regions of most violent earthquakes (Figs. 355 and 356) lie in belts that nearly coincide with the volcanic belts of the earth. They lie in a belt that circles the Pacific and also in one that joins the East and West Indies. A minor belt extends north and south through the Atlantic Ocean; another trends south from the major belt into Africa. The volcanic and violent earthquake belts in the main are regions also where mountains have been uplifted in late geologic time. The zones where the high segments and the low segments of the earth come together are the zones of weakness and the zones of greatest movement.

Not all earthquake centers are in the great deformation belts of the earth. Some are situated in areas remote from present-day volcanoes and late mountain building. New Madrid, Missouri, is far from volcanic centers and far from high mountains, and so is Charleston, South Carolina. Earthquakes are common in areas of surface loading as well as in areas of recent vulcanism.

Causes of Earthquakes.—Earthquakes, as stated, are caused by movements of the earth's crust which attend deformation by faulting

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[1] Milne, J., Earthquakes, pp. 1–388, 1913.

and folding. In general, the most severe shocks are in areas where volcanoes are active or where young mountains are formed. Volcanic eruptions commonly initiate seismic tremors. Earthquakes originate also in areas where adjustments of the surface are made necessary by gradation. At certain places the earth's surface loses weight owing to erosion; at other places it is overweighted by the deposition of sediments. Adjustments follow to meet new conditions. Where these take place suddenly, the movements are attended by shocks.

FIG. 355.—Earthquake map of the Atlantic Ocean. Stippling shows earthquake areas; close stippling, areas of violent earthquakes. (*From Heck.*)

Earthquakes as Geologic Agents.—Earthquakes are minor geologic phenomena that attend the deformation of the earth's crust. They are not major processes but are the results of major processes. Where the earthquake wave moves outward from the center of disturbance, it causes the earth to fissure, especially in loose material. Thus earthquakes which are caused by faulting may produce minor fissures and faults. Certain sandstone dikes have been attributed to the filling of fissures made by earthquakes. Often landslides attend earthquakes.

Fish and other organisms are doubtless killed by earthquake shocks, and their bodies probably accumulate in great numbers in certain strata, forming source beds for oil or fossiliferous deposits of various kinds. Earthquakes are important to man, because often they cause the loss of life and property.

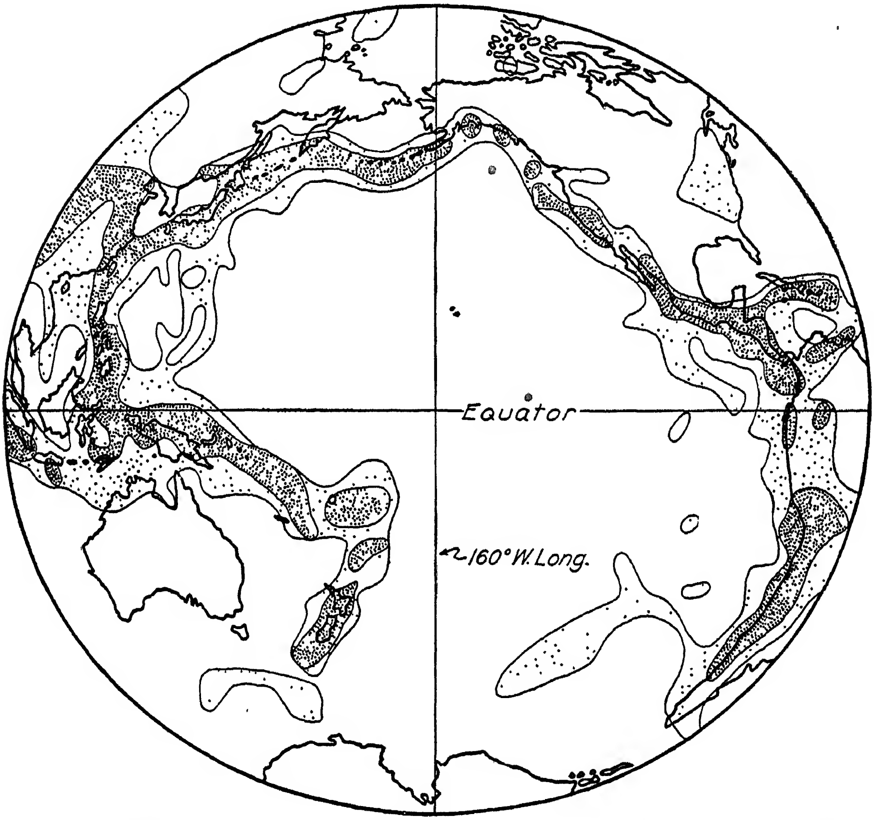

FIG. 356.—Earthquake map of the Pacific Ocean. Stippling shows earthquake areas; close stippling, areas of violent earthquakes. (*From Heck.*)

THE GREATER DIASTROPHIC MOVEMENTS

Diastrophism, as stated, includes all movements of parts of the solid earth with reference to each other. Evidences of such movements are numerous. Near the shores of certain seas, ancient shore lines are found back of the present ones, and sea cliffs cut by waves now lie beyond the reach of the highest waves (Figs. 357, 358). These show that the land has been elevated or that the ocean has withdrawn. If the ocean had withdrawn, the abandoned shore lines should be found at many places back of the present shores; but since the abandoned shore lines are found only in restricted areas, the conclusion is warranted that the land has been raised and that only these restricted areas have been

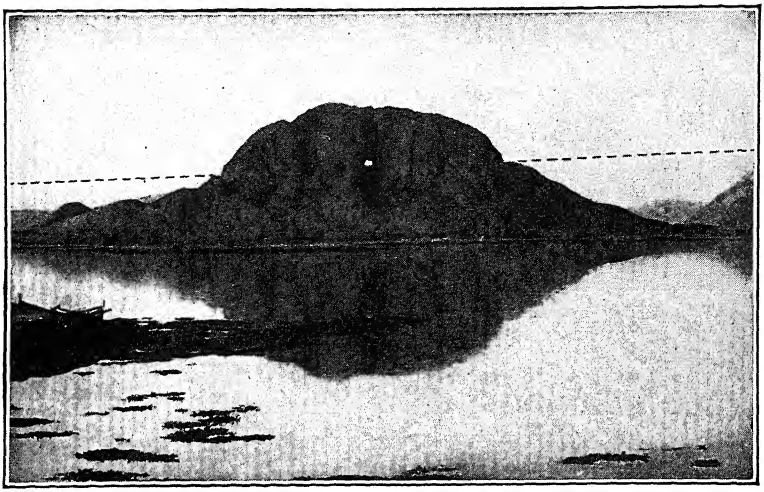

FIG. 357.—Photograph of Torghatten Island, west coast of Norway, showing wave-cut terraces and a wave-cut tunnel through an island of granite. The island is 860 feet above sea level. The wave-cut terrace, 400 feet above present high tides, is indicated by the dotted line. (*From The Changing World of the Ice Age by R. A. Daly, Yale University Press.*)

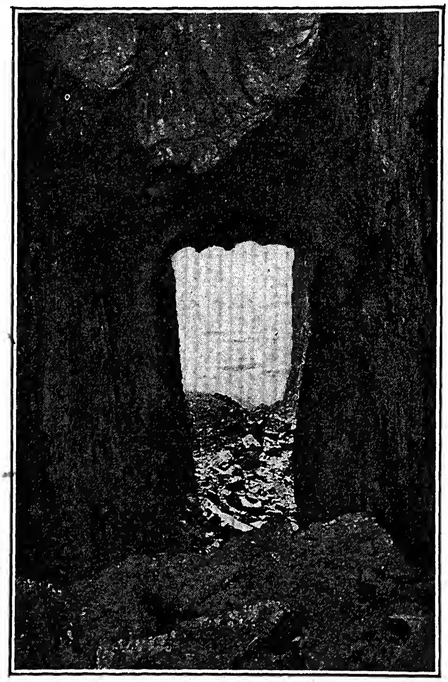

FIG. 358.—Closer view of tunnel shown in Fig. 357. The tunnel is about 90 feet wide, 245 feet high, 530 feet long. More than four million feet of granite was eroded out. (*From The Changing World of the Ice Age, by R. A. Daly, Yale University Press.*)

affected. Some of the ancient shore lines, moreover, are higher above the sea at some places than at others. The sea is essentially level. It is a little higher on shores near great mountain ranges where high rock masses attract the water by gravitative pull, but the differences in the level of the sea's surface are relatively small. The mean sea level is the most nearly constant level available, and as already stated it is taken as a datum for measuring the elevation of the land surface.

Rate of Movement.—The movements of the land are generally slow. Direct measurements have been made in countries bordering the Baltic Sea, by placing markers on its shores. It is found that the rate of elevation at places is 2 feet in 100 years. The northern coasts of Norway and Sweden have been raised 900 feet since the Glacial period, and the Labrador coast has been raised more than 400 feet (Fig. 359).

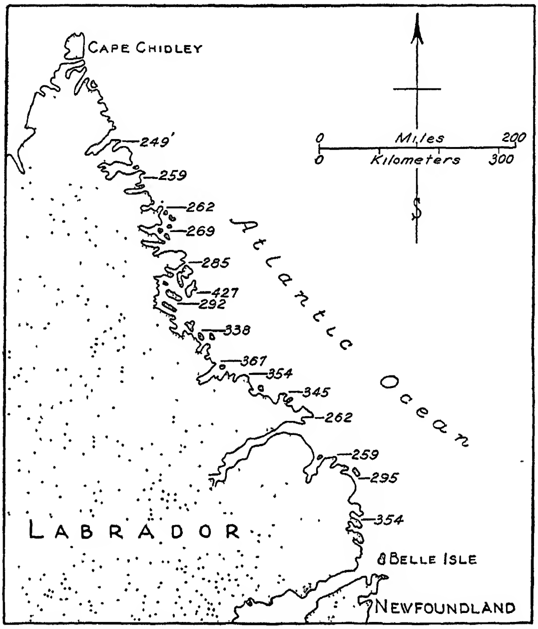

Fig. 359.—Map of the coast of Labrador showing the amount of post-glacial diastrophism as recorded by elevated marine beaches. Figures indicate the amount of uplift in feet above the present sea level. (*After Daly.*)

Not all movements, however, are slow. Displacements of the earth's surface during the San Francisco earthquake amounted to 21 feet, and during the great earthquake in Japan in 1923 the bottom of Sagami bay moved more than 1,000 feet. Both upward and downward movements are common. A classic record of movement is provided by the ancient Roman temple of Jupiter Serapis (Fig. 360) on the coast near

Naples, Italy. The columns have been bored by Lithophagus[1] 18 feet above the floor of the temple, and their shells are found in the holes. The temple was built by the early Romans; subsequently the land was submerged; and later it was raised above the sea.

"Drowned" valleys are common features along sea coasts, and submerged river channels extending far into the sea have been surveyed by soundings. These show that the land once stood at a higher level than it does today. Evidence of emergence is found far within the continents, where rocks containing the remains of marine shells are exposed far above the present level of the sea. Such rocks were deposited in the sea as essentially level beds. At places now they lie on edge, and the inference is warranted that they have been elevated and also deformed by diastrophism.

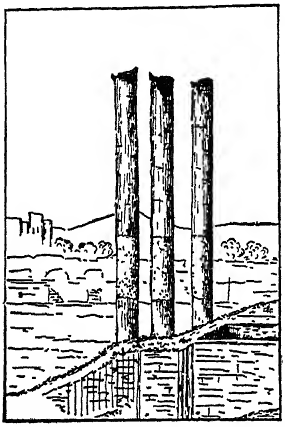

FIG. 360.—Ruins of the temple of Serapis near Naples, Italy. The temple was submerged and the stippled portions of the columns were bored by marine clams 18 feet above the floor of the temple. Subsequently the area emerged and the columns were raised to their present position above sea level. (*Data from Günther.*)

Kinds of Movements.—The greater movements of the earth are of two types: *orogenetic* and *epeirogenetic*. By orogenetic (mountain-making) movements the mountain masses are raised up. Nearly all great mountain ranges are anticlinoria; that is, the dominant folding is upward at the central parts of the folds. The formation of mountain folds generally is attended by strong horizontal movements along the circumference of the earth, which result in folding and in crumpling the strata.

By epeirogenetic[2] movements land masses of continental magnitude are raised and lowered with little folding. The study of the earth's strata and their relations to each other shows that the great crustal movements have not taken place continuously but that the movements occurred at intervals that were separated by periods of quiescence when the crust of the earth was undergoing only slight deformation. These great periods of deformation are called "revolutions." The folding of the strata to form the Appalachian Mountains is called the Appalachian Revolution, and the folding of the strata to form the Rocky Mountains is called the Laramide Revolution. These were periods of great mountain folding. Mountains were raised up at other times also,

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[1] Lithophagus is a genus of clams that perforate limestones and other substances in which they form flask-like excavations (Zittel).

[2] Orogenetic, Greek, *oros*, mountain; *gen.*, producing; epeirogenetic, Greek, *epeiros*, mainland, continent; *gen.*, producing.

and the most ancient rocks record very extensive earth movements. The great mountain movements are among the outstanding events of geologic history.

In addition to the great mountain-making movements which strongly deform the mountain belts and affect also the areas between the mountains, there are also epeirogenetic movements, or great warpings of the earth's crust that seem to be independent of the mountain-forming movements. These warpings are gentle but affect large areas. They raise segments of the earth's surface above the sea, where degradation prevails, and depress other segments below sea level, where aggradation is dominant. Thus the whole trend of geologic events is changed by such warpings (Figs. 361 to 363). In other regions plateaus are raised above land areas. These movements are radial, since they take place along the radii of the earth.

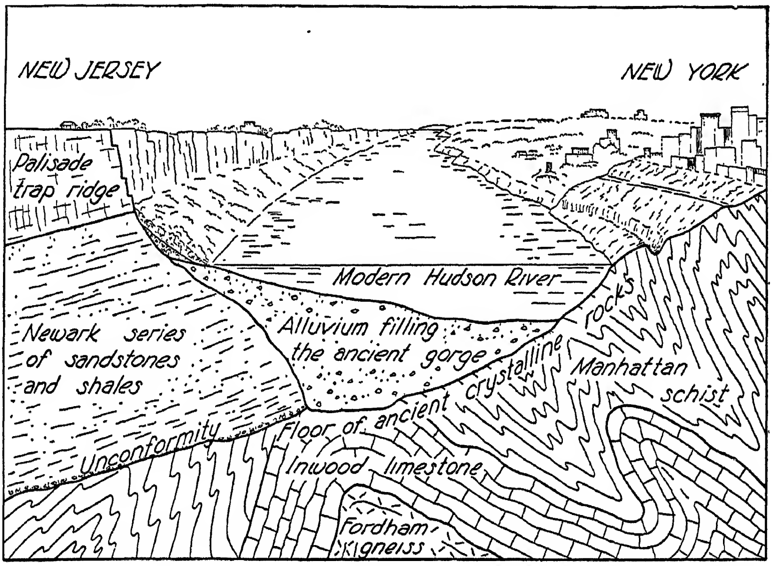

FIG. 361.—Generalized cross section of the valley of the Hudson River at New York City showing subsidence and filling of valley. (*Redrawn after Scott, courtesy C. P. Berkey.*)

TYPES OF DEFORMATION STRUCTURES

The deformation resulting from diastrophism may be classed as (1) *folds;* (2) *joints,* or fractures without appreciable displacements; and (3) fractures with displacements, or *faults.* These structures are treated on the following pages.

Attitude of Strata.—In order to show the position of an inclined surface such as a bedding plane in sedimentary rock, two observations are necessary: (1) the angle of slope of the strata and (2) the direction

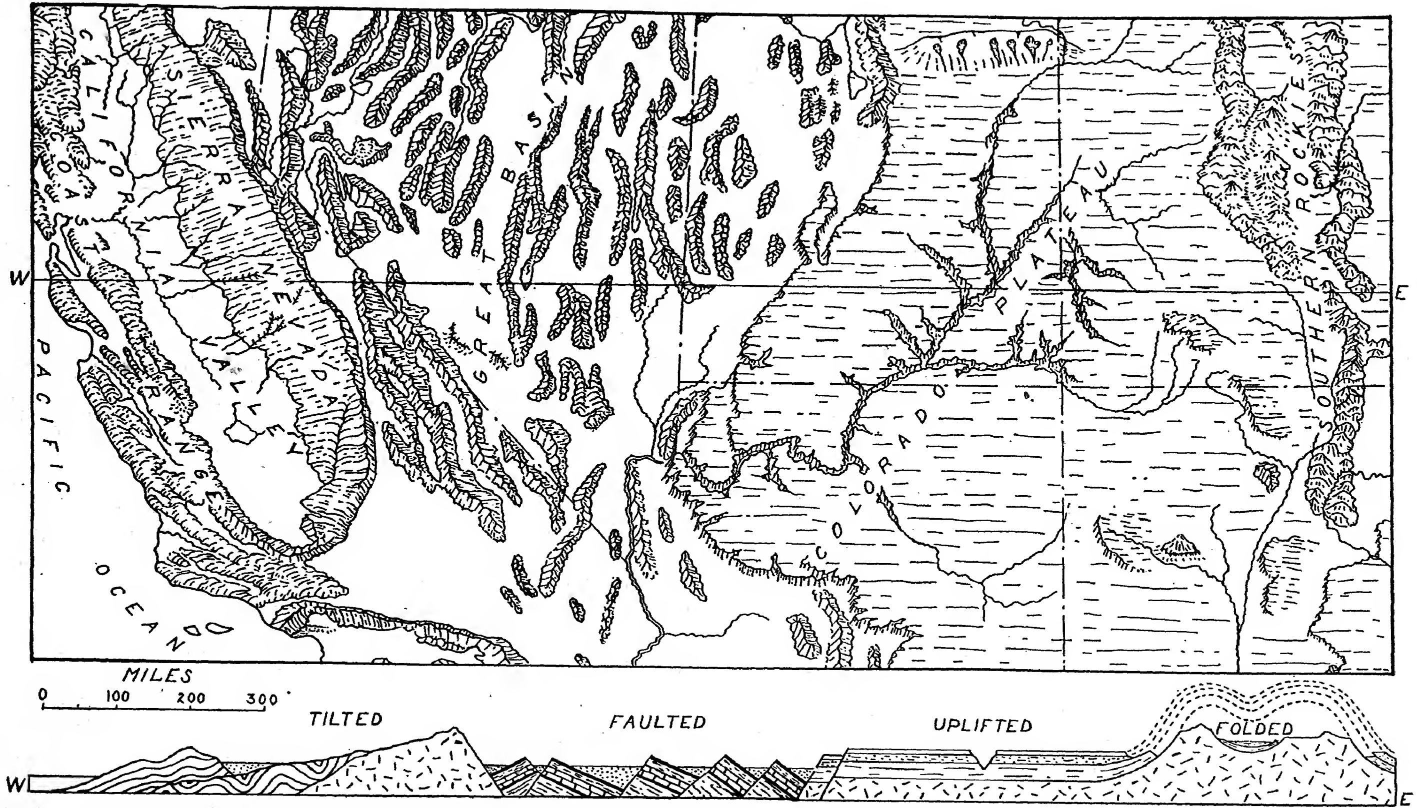

FIG. 362.—Physiographic map and generalized geologic cross section of the major physical divisions of southwestern United States, showing the influence of various types of diastrophism on the subsurface structure. (*Redrawn after Lobeck.*)

of the intersection of the strata with a horizontal plane. The surface of the earth may be considered to be a horizontal plane.

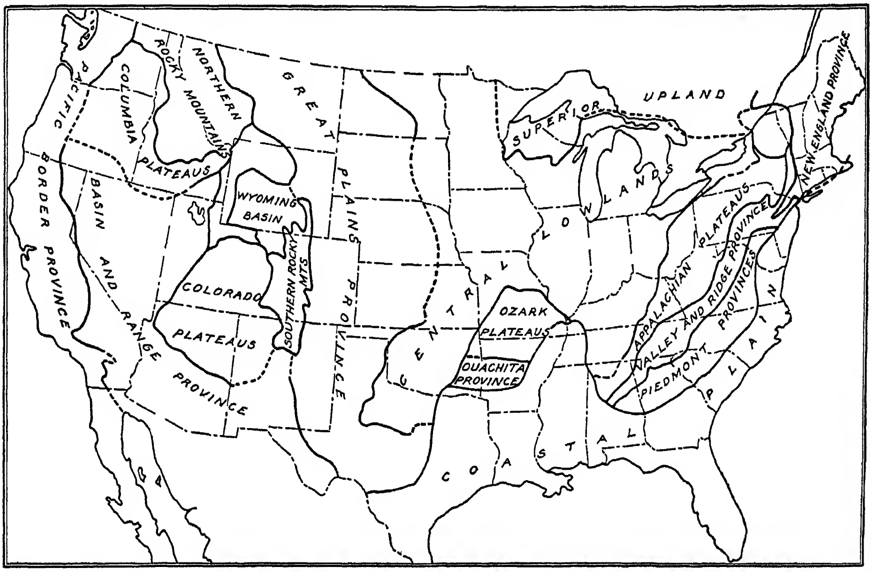

FIG. 363.—Major physical divisions of the United States. (*After N. M. Fenneman and U. S. Geol. Survey.*)

The term *dip* is used to designate the angle of inclination or the amount a bed is tilted from the horizontal position (Fig. 364). The dip is an angle in a vertical plane and is measured downward from the horizontal direction.

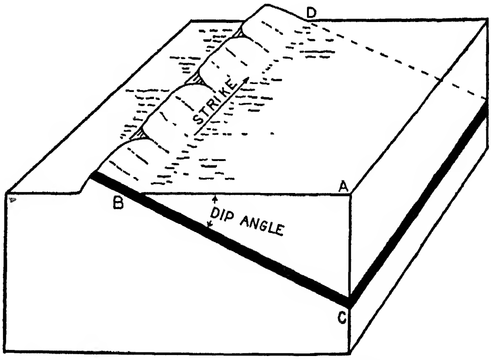

FIG. 364.—Diagram illustrating the dip and the strike of a tilted stratum. *ABC* is the angle of dip. *BD*, a horizontal line, is the direction of strike. The direction of dip is toward the right at right angles to the direction of strike.

The term *strike* designates the direction of the intersection of a stratum with a horizontal plane (Fig. 364). The direction of strike is measured by means of a compass (Fig. 365) the dial of which is graduated to degrees. Thus if the line of intersection of the bed and a horizontal plane (*BD*, Fig. 364) extends in a direction 40 degrees east of north, the strike is recorded as N.40° E. The direction of dip is always measured at right angles to the strike.

The compass (Fig. 365) used for geological mapping usually contains a clinometer also, which is either a pendulum or a mounted level. Thus the same instrument may be used for determining the direction of strike and the amount of the inclination, or angle of dip. Dip and strike together define the position, or *attitude,* of a stratum with respect to a horizontal surface and to compass directions. Where beds dip in a single direction, they form a structural feature that is called a homocline (Fig. 366). Very commonly the term monocline is used to define the same feature, although as originally used it describes a fold or flexure on both sides of which the rocks are essentially level (Fig. 367).

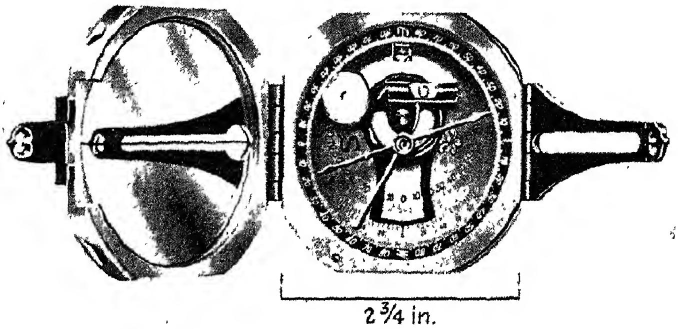

Fig. 365.—Brunton compass.

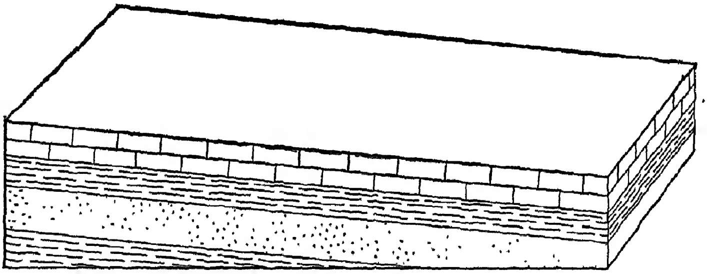

Fig. 366.—A homocline; the rocks dip in one direction at about the same angle.

Folds.—In many regions the sedimentary rocks that were deposited originally as nearly horizontal strata are found buckled into more or less symmetrical series of folds with alternating crests and troughs. Two terms are commonly used in describing such folds. Where strata are arched up as in the crest of an upfold, they form an *anticline* (Fig. 368);

the trough, or downfold, between crests is a *syncline* (Fig. 368). Each of these may have various modifications. An anticline may coincide with a hill and a syncline with a valley, but after erosion the anticlines commonly lie in the valleys and the synclines form the hills. The axial plane of a fold is the plane that may be considered to pass through the center of the fold (Fig. 369). The axis is the line which the intersection of the axial plane makes with the bedding planes of the folded series.[1] If the axis is inclined, the fold is a plunging fold (Fig. 370). The two sides of a fold are called its limbs. If the limbs dip at about the same

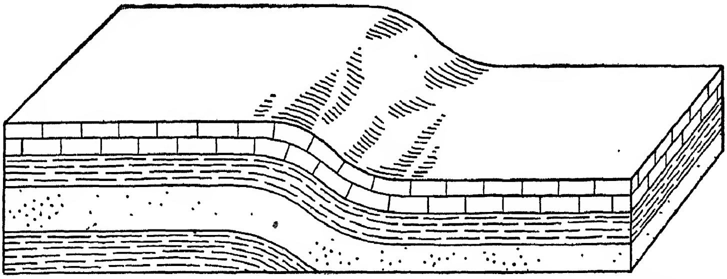

FIG. 367.—A monoclinal fold. The rocks dip in the same direction at different angles. On either side of the monocline the strata are either flat-lying or have uniform low dips.

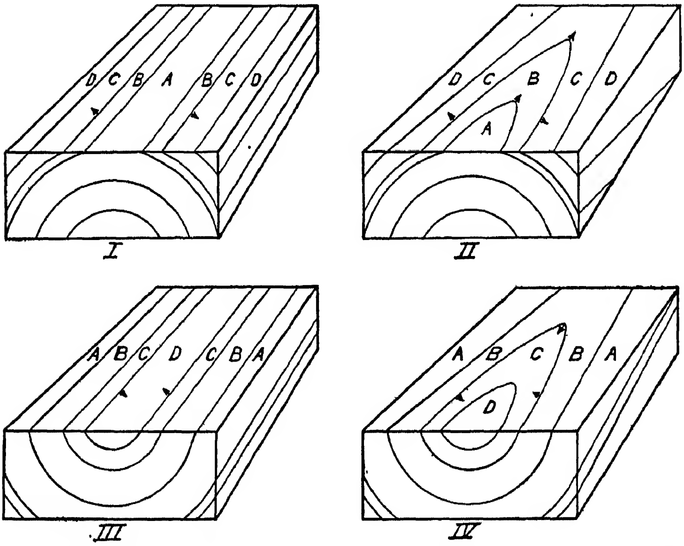

FIG. 368.—Diagrams showing the relative age and structural relations of strata included in simple folds. *A* is the oldest stratum that crops out at the surface, *D* is the youngest. *I*, symmetrical anticline; *II*, pitching anticline; *III*, symmetrical syncline; *IV*, pitching syncline.

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[1] In mapping it is common practice to map the intersection of the axial plane and the surface of the earth as the axis of the fold—that is, to map the line from which the rocks dip in both directions.

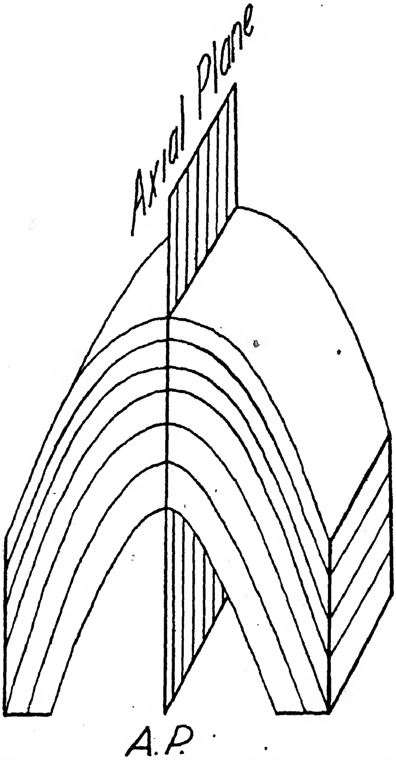

FIG. 369.—Sketch showing the axial plane of an anticlinal fold.

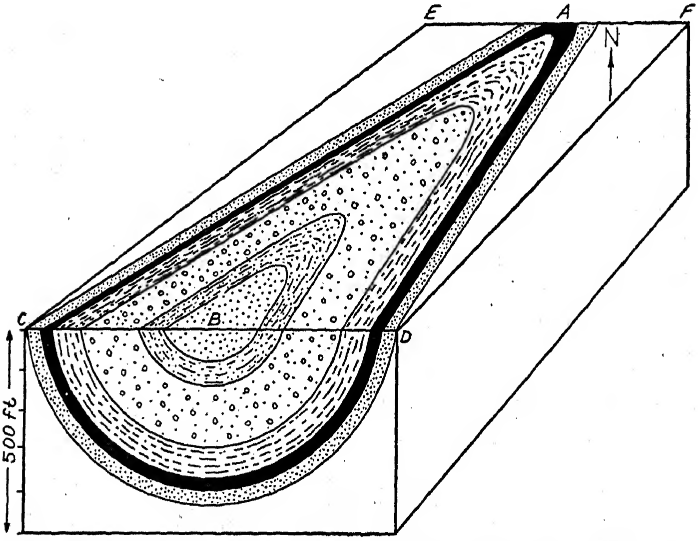

FIG. 370.—Block diagram of a syncline plunging toward the southwest. *ECDF* represents the plane of the earth's surface. At *A* the formation shown in black is at the surface but at *B*, due to the plunging of the fold, the same formation is more than 300 feet below the surface.

angle, the fold is said to be symmetrical; and if one dips at a higher angle than the other, the fold is asymmetrical. An overturned fold is one in which one limb is at places doubled under, so that it lies below the other (Fig. 371). A recumbent fold is an overturned fold in which the limbs are essentially horizontal (Fig. 372). An isoclinal fold is one in which the two limbs dip equally in the same direction (Fig. 373).

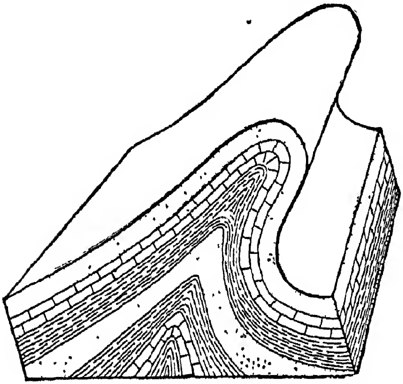

Fig. 371.—An overturned fold.

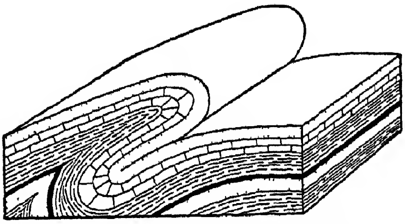

Fig. 372.—A recumbent fold.

In nearly all folds the axis is not a level line but plunges or pitches from the horizontal position (Fig. 370). If it plunges at both ends, it is called a double-plunging anticline. A dome is a structural feature in which the rocks dip away from a common center in all directions. The double-plunging anticline is really an elongated dome, and in actual field practice it is generally called a dome. If the axis of the anticline rises into the air, it is called an open anticline. If it does not but plunges at both ends, it forms a "dome," or in oil-field practice a "closed structure" or "closed anticline." Many structural features are so small that the structure is seen at a glance; and in arid mountain regions where vegetation is scarce, large structural features may be seen with little difficulty. In general, however, the larger structural features are discovered only by mapping the area. Observations are recorded on a map, and the attitudes of beds and faults and contacts between rocks are shown by symbols (Fig. 374).

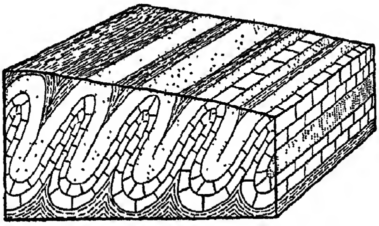

Fig. 373.—Sedimentary beds closely folded to form isoclinal folds.

Refolded Folds.—In many regions the earth's surface has been folded more than once along the same axes. Figure 375 shows an anticline that has been eroded to a plain. Later the area was submerged, and the folded beds were covered by later sediments. Subsequently the area was folded again; and since the ancient axis was a plane of weakness, the beds yielded along it a second time. A cross section of the area shows the beds at the surface dipping gently from the axis and the lower beds dipping away from the axis at higher angles.

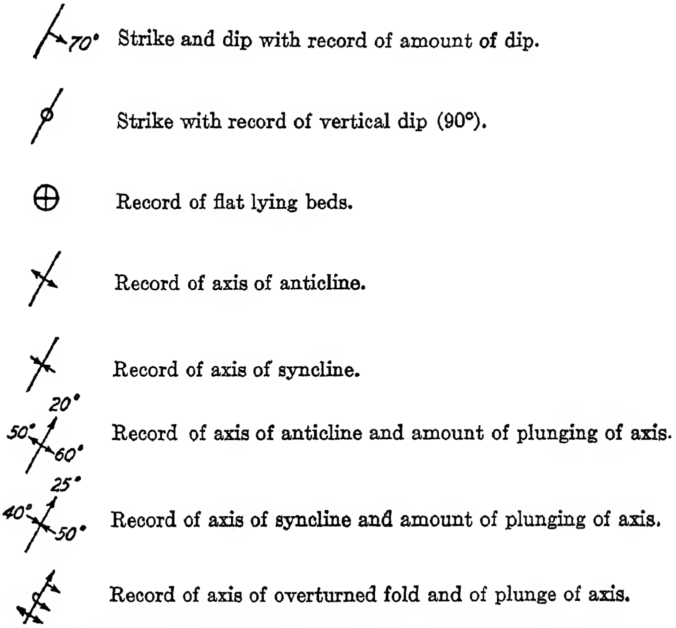

Fig. 374.—Conventional signs used on geological maps.

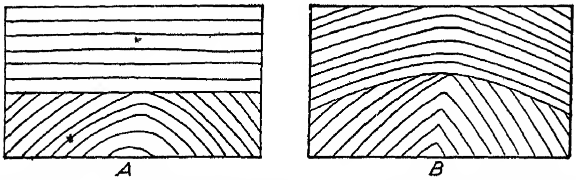

Fig. 375.—Cross section of a refolded anticline. *A*, arched beds eroded and covered by flat-lying beds; *B*, the same after folding along the same axis.

Differential Compaction.—The folds of the earth's surface, in general, are due to compressive stresses acting along the earth's circumference. Certain minor folds, however, are believed to have been formed by the settling of rocks above an ancient irregular surface. Figure 376*A* shows a series of shales and sandstones deposited above a rigid hill. If, as a result of pressure, water is squeezed out of the shales and the shale shrinks 20 per cent, the overlying rocks will be let down above the ancient hill, and the rocks away from the summit of the hill will be let down more because the shales are thicker than at the summit. The sandstone (Fig. 376*B*) will dip away from the center of the hill, forming an anticline. According to certain investigators, some of the minor structural features in the oil fields of Kansas and Oklahoma have formed by this process.

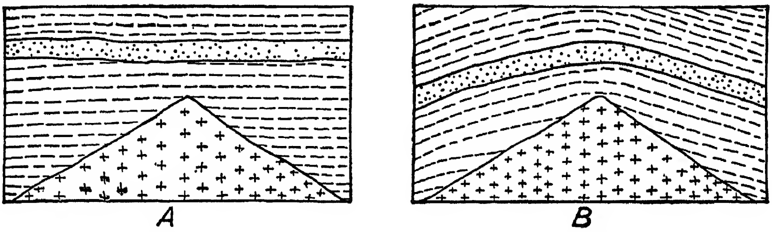

FIG. 376.—Diagrams illustrating an anticline formed by differential compaction. *A*, before compaction; *B*, after compaction of mud or shale by shrinking 20 per cent.

Systems of Folds.—A large system of folds, including both anticlines and synclines, in which the dominant folding is upward is an *anticlinorium* (Fig. 377). A *synclinorium* is a system in which the dominant folding is downward. A *geanticline* is an anticline or an anticlinorium of very large dimensions, and a *geosyncline* is a syncline or synclinorium of very large dimensions.

The areas of the great mountain folds are characterized by uplifts that extend for hundreds and even thousands of miles. Axes of individual

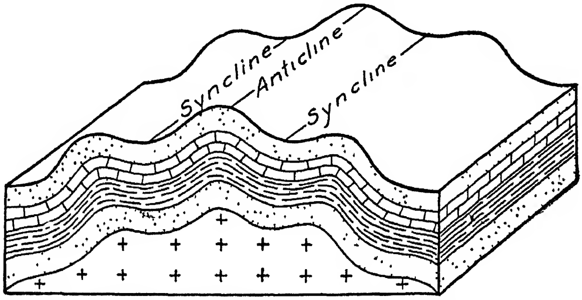

FIG. 377.—Sketch showing an anticlinorium.

folds may be traced for 100 miles or more. In anticlines the beds dip or slope away from the axes. In synclines the beds dip toward the axes. Generally the axes are not level but plunge at places, and that gives the impression of folding of the axes. The axes of the major anticlines of a mountain region overlap each other, and small anticlines or domes are developed on the flanks of larger ones. At places prominent cross folds are developed. The spacing of the folds and the undulations of the axes of major folds show that the earth's surface has been shortened in two directions—that is, at right angles to the trend of the axes of the major folds and also in the directions of the major axes. In most mountain regions where close studies of folds have been made, however, shortening across the trend or axes of major folds is several times that which has taken place along major axes.

During the formation of great mountain ranges by folding, the areas between the mountains also are affected. In the great downfolds, or geosynclines, between the mountain ranges, the rocks have a general dip away from one mountain range, and they rise toward another.

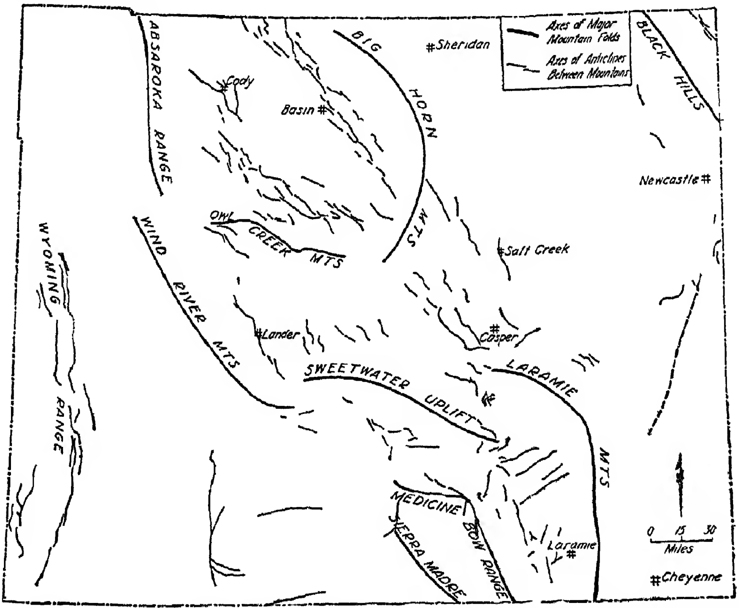

FIG. 378.—Sketch of Wyoming showing axes of major mountain folds (heavy black lines) and intermontane minor folds (light black lines). (*After Hares, Heald, Richardson, Woodruff, Collier, and others, U. S. Geol. Survey. Wyoming Geol. Survey.*)

Between the areas of great mountain folds are smaller folds, and since many of these have axes that are parallel to the greater folds of the neighboring mountains, it is believed that they were formed at about the same time and that they are the results of the same forces. The arrangement of the minor folds in Wyoming is illustrated by Fig. 378, which shows the positions of the major mountain uplifts and of the smaller anticlinal folds around them. Essentially the area of the entire state has been involved in the uplifts that formed the Black Hills, Bighorn Mountains, Shoshone Mountains, the Laramie, and other mountain ranges of Wyoming.

Structure-contour Maps.—Folded beds commonly lie one above the other like cards in a flexed pack. If an upper one which may be seen at the surface is folded, lower ones also will be folded. If the rocks of an area are known from drilling them or by measuring their thickness in

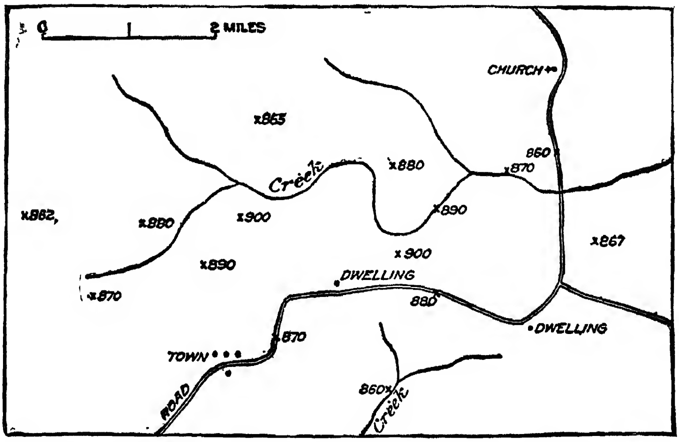

FIG. 379.—Sketch map showing elevation of the same stratum at different points, marked by crosses. (*After Gardner.*)

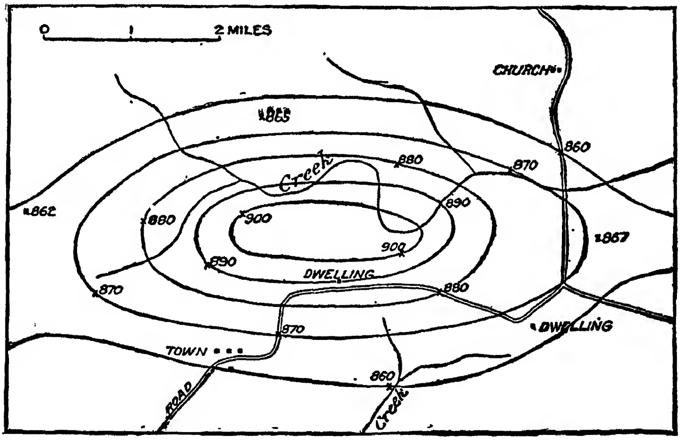

FIG. 380.—Sketch map showing elevations of same stratum at different points marked by crosses as in Fig. 379. The structure contours connect points of equal elevation, thus outlining an elongated dome. (*After Gardner.*)

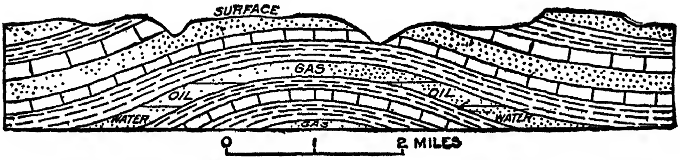

FIG. 381.—Lengthwise section of elongated dome shown in Fig. 380, vertical scale greatly exaggerated. (*After Gardner.*)

exposures that are located in the general region, it is possible to map the structure of a bed by determining various points where it is seen or where it can be calculated to lie below the surface (Fig. 379). If rocks above the bed or datum plane which is being contoured are exposed, the depth of the bed may be calculated. If the rocks exposed are those that lie below the bed that is contoured, then the position of the latter before it was removed by erosion may be determined. By utilizing these data one may contour the bed and show on a map by structure contours its approximate position at all places over the entire area mapped (Fig. 380). Since the other beds lie parallel to it, the contour map will show the positions of all beds of the series. A section through the area contoured is shown by Fig. 381. Since most of the world's petroleum is found in domes, the mapping of oil-bearing areas by structural contours is a great aid to the search for petroleum. Contour maps showing structure are used also in the explorations for coal, iron ore, phosphate rock, and other valuable beds.

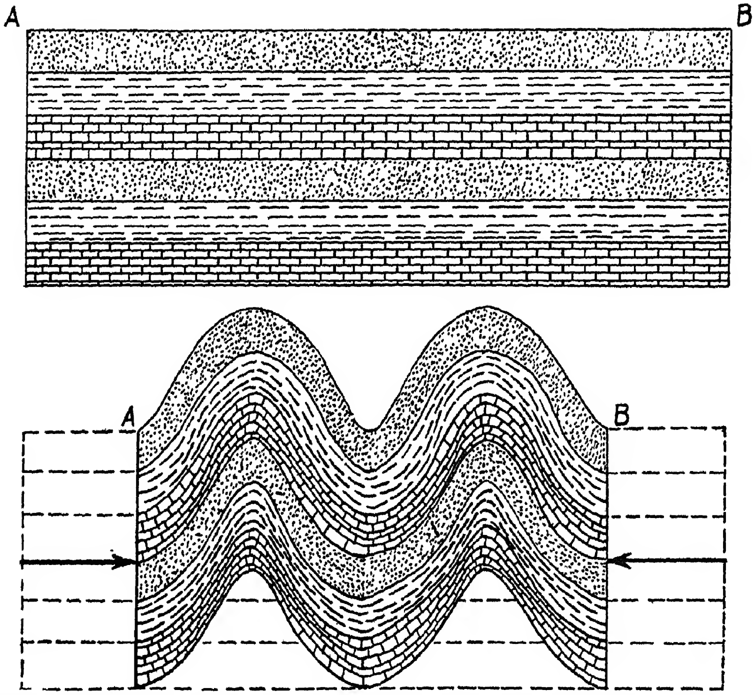

Fig. 382.—The upper figure shows a system of beds horizontal as laid down. The lower figure shows same after folding. *AB* in lower figure is two-thirds as long as *AB* in upper figure. The shortening is one-third.

Shortening of the Earth's Crust.—The processes of folding result in shortening the earth's crust, as is illustrated by Fig. 382, where the bed represented by *AB* is 6 miles long and the distance after folding is 4 miles. The length of the area formerly covered by the bed *AB* has decreased one-third by folding. The shortening shown by Fig. 383 is 10 per cent. It is estimated by R. T. Chamberlin that the shortening of the earth's circumference by the crustal folding attending the formation of the Appalachian Mountains near Harrisburg, Pennsylvania, including

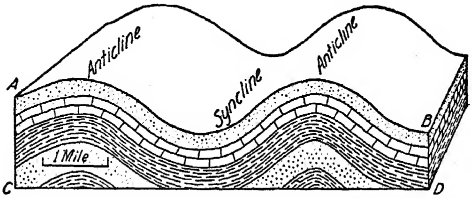

FIG. 383.—Diagram showing anticlines and synclines of folded beds. The curved line *AB* represents a line 7 miles long. The distance *CD* is 6.3 miles long. By folding, the distance *AB* has been shortened about 10 per cent.

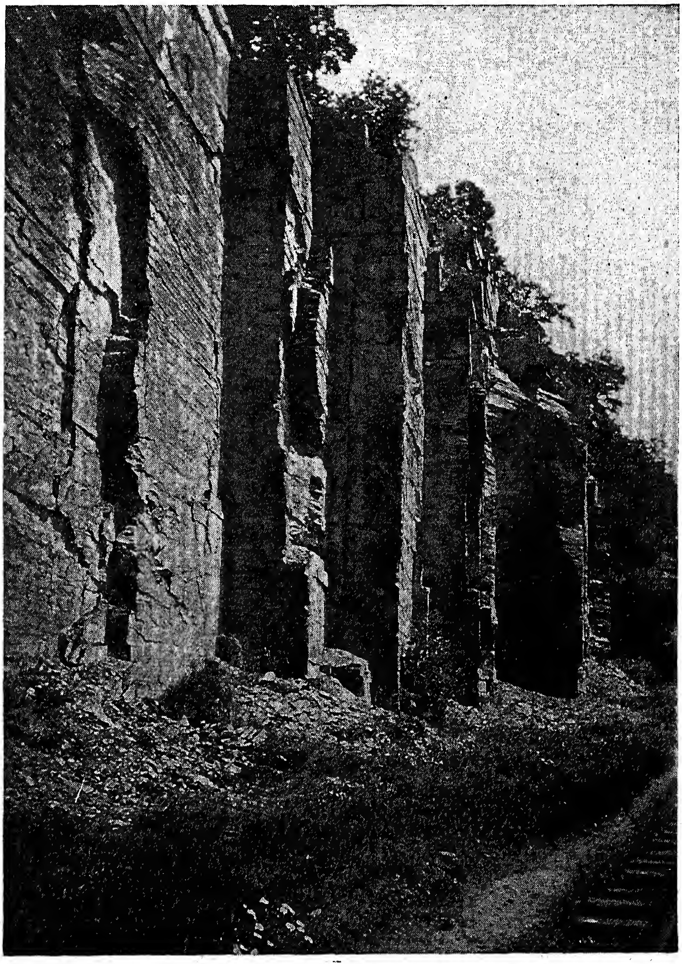

FIG. 384.—Vertical jointing in Portage beds, Cayuga Lake, New York. (*Photograph by Kindle, U. S. Geol. Survey.*)

the crystalline belt, has amounted to 100 miles; and according to Heim, the shortening due to the folding of the Alps has resulted in shortening the earth's circumference about 125 miles. In the Rocky Mountains of North America, in an area north of Denver in Colorado, the folding of the mountains, according to Chamberlin, appears to have shortened the earth's crust about 8 miles. At places rocks are thrown into comparatively open folds, as shown by Fig. 383; at other places they are closely folded (Fig. 373).

Joints.—Nearly all consolidated rocks are crossed by deep cracks and fissures (Fig. 384). Some are no wider than a sheet of paper, and others are so thin that there is no appreciable opening, yet the fissure is revealed when the rock is weathered or when broken with a hammer. Still other fissures show narrow gaps. A joint is a fissure along which there has been no appreciable movement. Most rocks are cut by three or more systems of joints, and they may be seen nearly everywhere that rocks are exposed. Joint planes are rarely followed more than a few score feet, though some may extend for hundreds of feet. In certain rocks joints are closely spaced so that the rocks are broken into small blocks. In some, however, the joints are widely and regularly spaced, and large blocks may be broken out and used for monoliths in buildings.

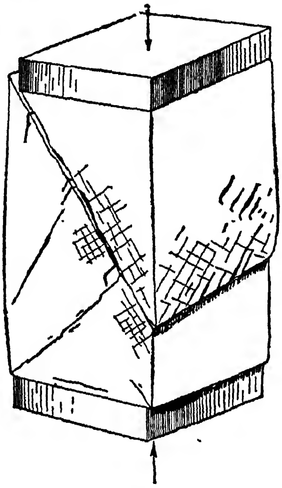

Fig. 385.—Block of material deformed by pressure applied at ends. (*After Daubrée.*)

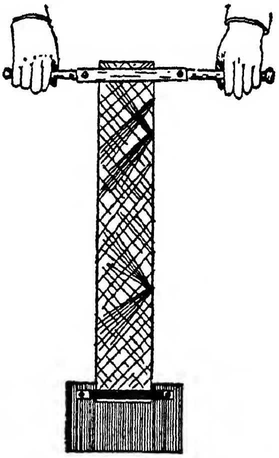

Fig. 386.—Sketch showing effect of torsion on glass plate. Lower end was held fast to block, and upper end was twisted. (*After Daubrée.*)

Joints are formed by stresses attending the deformation of the outer part of the earth. Nearly all of the earth's surface has undergone stress, for small folds are found in the great areas between the mountain uplifts in nearly all of the great geosynclines between the mountains. All but the very young rocks are appreciably jointed.

Many joints are the results of compressive stresses. A block of granite under compression develops a system of joints nearly at right angles to each other, as is illustrated by Fig. 385. Torsion also develops joints, as shown by Fig. 386. Joints likewise are formed by tensional stresses or as cooling cracks, such as are seen in many intrusive rocks and lava flows.

Fissures.—A fissure is an opening or parting in the rocks. It may be a mere crack with no visible open space, or it may present a wide-open space. Movement may have taken place at right angles to the plane of the crack, or it may have taken place along it. Where the movement has taken place along it, the fissure is a *fault.* Fissures include joints and faults, but many fissures are larger and more persistent than joints. Some fissures may be traced many miles, and fissure systems are followed scores or even hundreds of miles. A sheeted zone is a mass of rock cut by closely spaced parallel fissures. If there has been shearing, it is a shear zone. A fractured zone is a mass of rock cut by a large number of small, irregular fractures, the mass as a whole being more or less tabular. Fissures have great economic importance, because metal-bearing veins are deposited in and along them. Fissures, like joints, are formed, in the main, by compressive stresses, although some may be formed by torsional or by tensional stresses.

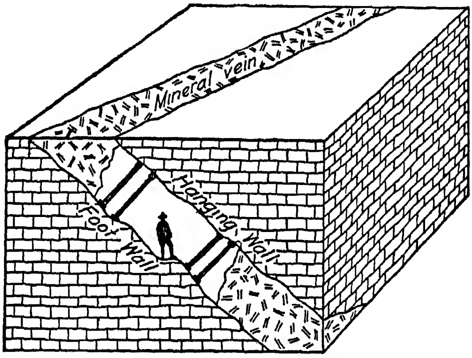

Fig. 387.—Diagram showing a mineral vein in flat-lying limestone. The miner stands on the footwall, and the hanging wall "hangs" above his head.

Faults.—A fault is a fissure along which there has been displacement. Any fissure is accompanied by some movement; otherwise there would not be a break in the rock. Most faults are inclined; a few are vertical. The rock on the sides of the fault is the wall rock. The footwall of the fault is the rock below; and the hanging wall, the rock above the fault. These are mining terms and are applied to both faults and veins. The footwall is the rock on which the miner stands as he works a vein, and the hanging wall is the rock above his head (Fig. 387). Cross sections including faults are shown by Fig. 388.

By movement along faults rocks are broken and ground up, and the fault fissure is commonly filled with the broken material, *drag* (Fig. 389). If such material is made up of distinct fragments, it is *fault breccia.* If it is finely ground, it is *gouge.* In many faults this material is so fine that it is clay-like in consistency; such material is *clay gouge.* The fragments in a fault fissure are made up of the rocks that the fault crosses, and often the fragmental material is a mixture of many kinds of rocks. At places the wall rocks along faults are polished by movement. Such polished surfaces are called *slickensides.* Fault *striae* are scratches on the walls of faults formed by the abrasion of hard particles against the walls. On some faults two sets of striae cross, showing different movements at different times. Fault *grooves* are furrows that are deeper than striae but are similarly formed. Because usually they record movements by strong forces, they have greater significance than striae.

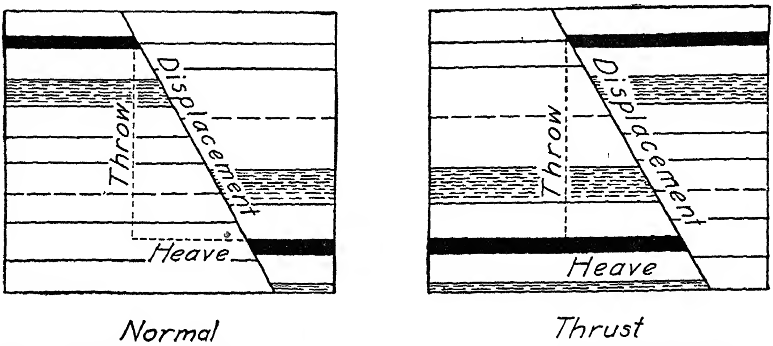

FIG. 388.—Diagrams showing cross sections of a normal fault and of a thrust fault. The figures illustrate displacement, throw, and heave.

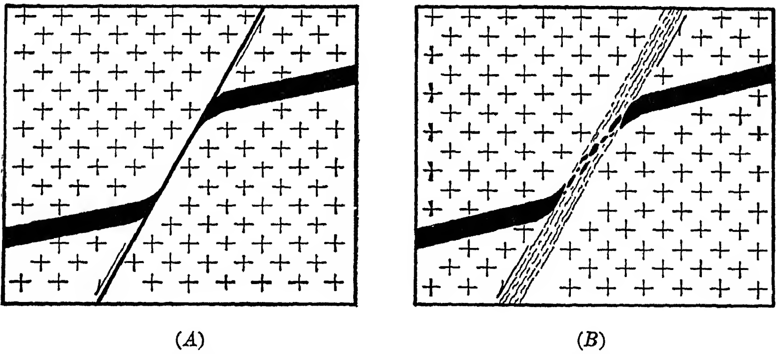

FIG. 389.—*A*, cross section of a vein curved near a normal fault; *B*, cross section of a vein curved near a normal fault with broken fragments of the vein in the fault gouge between segments of the vein.

The walls of faults have moved up or down or horizontally. In actual practice it is generally not possible to ascertain which wall has moved and the exact direction of movement. Cross sections of the faulted areas are constructed; and if on these the hanging-wall block appears to have dropped, the fault is called a *normal fault;* if it appears to have risen, the fault is a *thrust,* or *reverse fault* (Figs. 388, 390). This practice is necessary, because the same relations appear whether the hanging wall moved down or the footwall moved up.

The block that on the section appears to have moved up is the upthrow side of the fault, and that which appears to have moved down is the downthrow side. The striae and grooves on the fault walls do not record the direction of movements. They show only that the movement has taken place in one of two directions, and they do not show which block has moved. On some fault planes there are raised places which are due to hard spots in the rocks. They are worn deeper in the direction from which the movement came. Some slickensided surfaces show therefore that movement has been in one of two directions. When the hand is passed over it, the fault surface will feel smoother where the movement of the hand coincides with the movement of the abrading material than when the hand is moved in the opposite direction (Fig. 196).


FIG. 390.—Horizontal thrust showing slickensides on base of marble at Burlington, Vermont. (*Photograph by Macan.*)

Although it is often impossible to determine the extent of movement along a fault, the lost segment of a faulted bed or vein often may be discovered by mapping the area. The attitude of a fault is defined by its strike and dip. The hade of the fault is its inclination to a vertical plane. It is the complement of the dip.

A *fault scarp* is a cliff or a steep slope developed on the upthrown block of a fault. Often erosion has reduced the upthrow block to the same level as that of the downthrow block. Where faults cross bedded rocks, the beds commonly are displaced by the faulting, owing to drag. In Fig. 389 the pressure of the overlying block has bent the vein near the fault plane. Normal faulting appears to extend the earth's crust in the region faulted, and reverse faults appear to shorten the region. The bed *CD* shown in Fig. 391 was 33,600 feet long before faulting; after

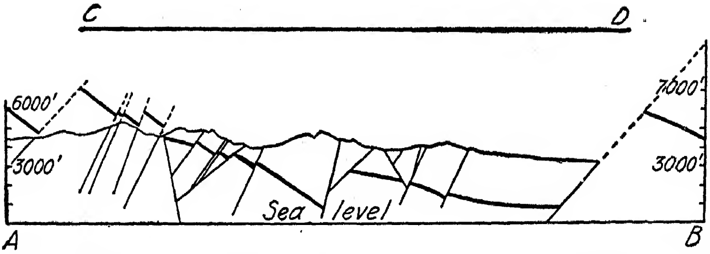

FIG. 391.—Cross section through a faulted area near Rhyolite, Nevada. The heavy short lines represent segments of a faulted rhyolite flow. At left of figure this flow is eroded, but its position before erosion may be estimated because the positions of lower flows are known. Laid flat, end to end, the segments of the flow would make a line *CD*, 33,600 feet long. The horizontal distance over which the sections are now distributed is 42,240 feet. The difference is 8,640 feet or 25.7 per cent of 33,600 feet. (*After Ransome, Emmons, and Garrey, U. S. Geol. Survey.*)

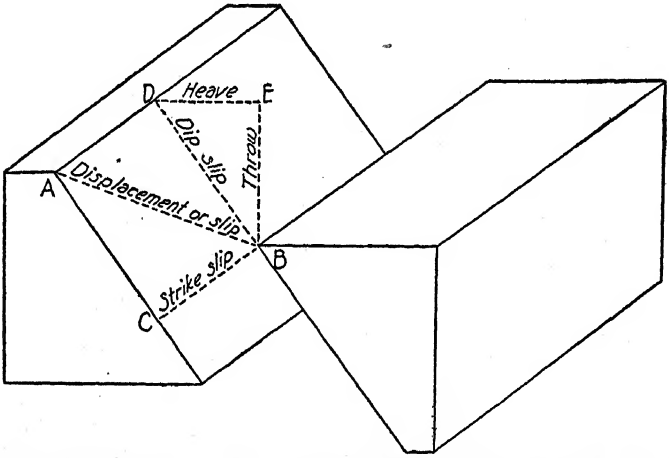

FIG. 392.—Faulted block showing elements of faulting. *BDE* is at right angles to line *AD*.

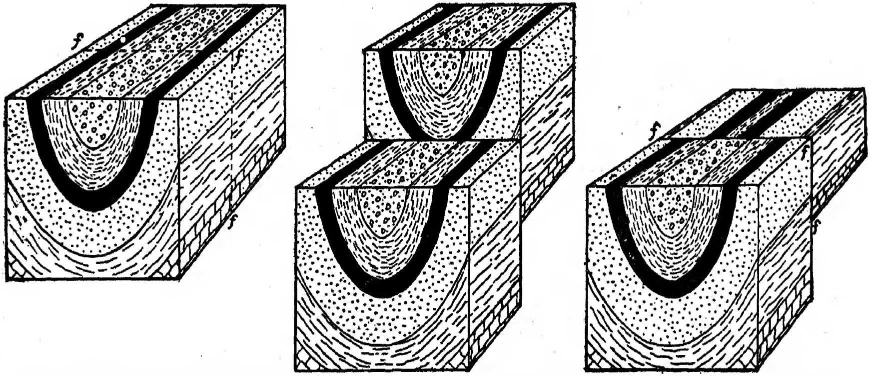

FIG. 393.—Diagrams illustrating a dip fault cutting a syncline. Faulting was followed by peneplanation.

faulting it seems to be extended with interruptions over an area *AB*, or 42,240 feet long. This may be interpreted as showing that the area has been extended 25.7 per cent, but not all fault blocks move directly down the dip, and there are many corrections to be made in interpreting the extensions produced by faulting.

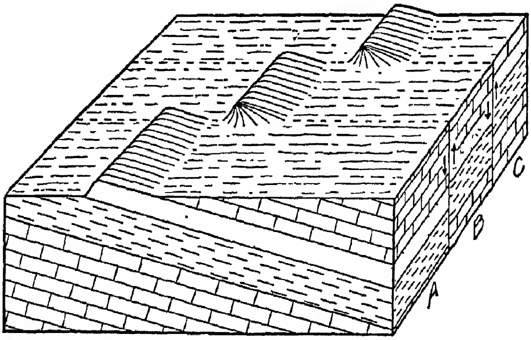

FIG. 394.—Diagram showing the effects of faulting in an area with tilted beds of unequal hardness. The faults are vertical dip faults. Block *B* moved upward with relation to *A*. Block *C* moved upward with relation to block *B*. There was no horizontal displacement. The faulting was followed by erosion and hogback ridges remain. The apparent displacement of the ridge is not due to horizontal movement, but it is the result of monoclinal shifting of the outcrop during the erosion of the vertically elevated blocks. (Compare with Fig. 396.)

Elements of Faults.—In working out fault problems, particularly in connection with the search for faulted segments of rocks or ore bodies, it is good practice to indicate the components of faulting. The displacement (Fig. 392), called also the slip, is the separation as measured on the fault plane; the throw is the vertical displacement; and the heave is the horizontal displacement. The slip, throw, and heave form a right-angle triangle.

Faults that cut across the dip of the beds and lie at right angles to the strike are *dip faults* (Figs. 393, 394); those that lie parallel to the strike of the beds but cross the dip are *strike faults* (Fig. 395); and those that cross both dip and strike are *oblique faults* (Fig. 396). By normal faulting, certain beds may be cut out of the series where it crops out at the surface, as is shown by Fig. 397, where bed 7 lies against bed 9 and beds below 7 are concealed. Beds may be cut out also by reverse faulting, as is shown by Fig. 398, where beds 4 and 7 are adjacent and beds 5 and 6 are cut out by erosion. In bedding-plane faults the fault blocks move parallel to the beds (Fig. 399).

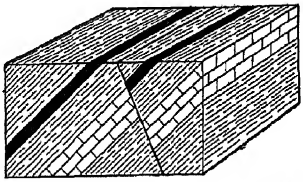

FIG. 395.—Diagram showing sedimentary beds displaced by a normal strike fault. Erosion has planed the area so that the faulted blocks are at the same elevation.

An anticline may pass into a fault (Fig. 400). A block depressed between two faults is a *graben* (Fig. 401), and a block raised between two faults is a *horst* (Fig. 402). Faulting and tilting of faulted blocks may go on together and thus parallel ridges, or "saw-tooth" mountains may form. The Bullfrog Mountains near Rhyolite, Nevada, are a striking example of a group of mountains outlined by faults (Fig. 391).

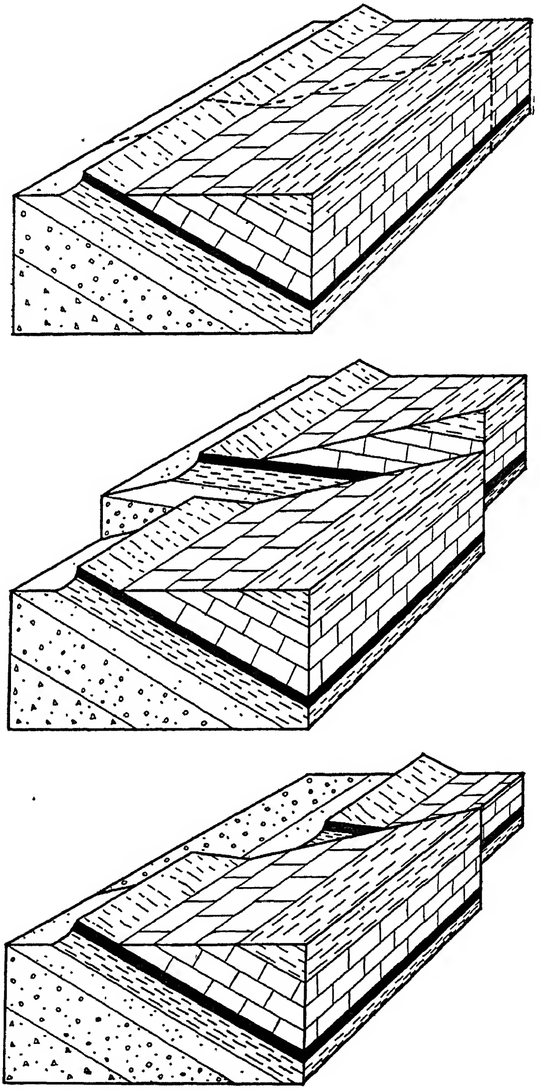

Fig. 396.—Diagrams showing an oblique fault cutting inclined strata. Faulting was followed by erosion.

Reverse Faults.—A reverse fault is one where the hanging wall ppears to have been raised or the footwall depressed. The fault plane ips toward the raised block. Reverse faulting, like folding, results in

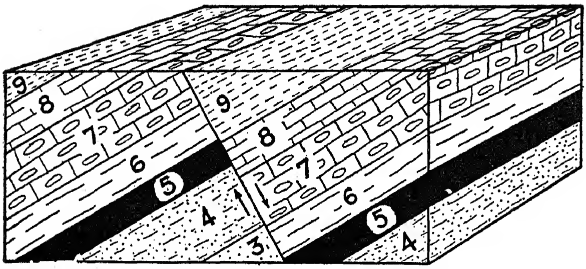

Fig. 397.—Diagram showing how beds may be concealed by normal faulting followed by erosion.

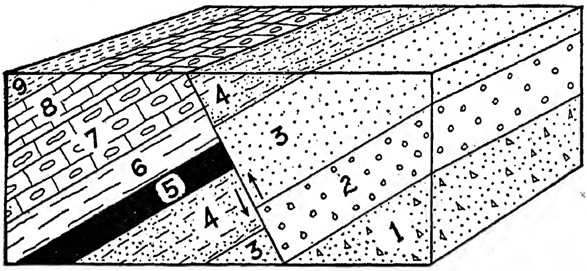

Fig. 398.—Diagram showing how beds may be cut out by reverse faulting followed by erosion.

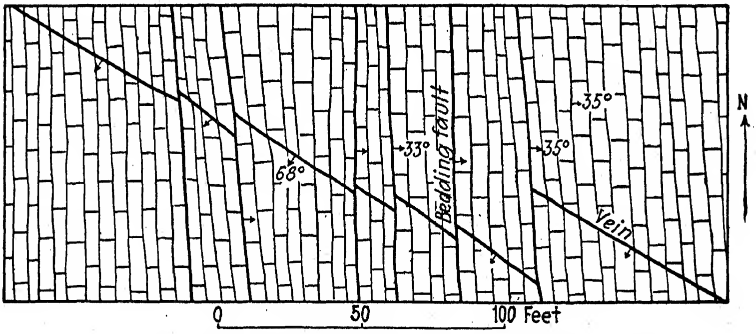

Fig. 399.—Plan of Headlight vein tunnel level, Philipsburg, Montana. The vein cuts across the bedding of the country rock and is displaced by faults which follow the bedding planes.

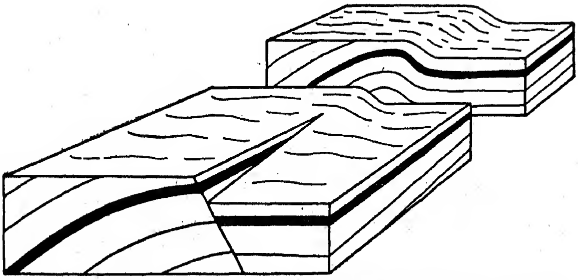

Fig. 400.—Diagram showing an anticline passing into a fault.

shortening the earth's crust. In the main, reverse faults dip at low angles, and such reverse faults are called also "low-angle faults." In general, these faults dip from 0 to 45 degrees, although some are steeper. Many low-angle faults are ruptured folds broken along an axis of folding (Figs. 403, 414).

Great thrust faults are found at many places. Some are traced scores of miles and have horizontal displacements of more than 20 miles. The Lewis overthrust (Chief Mountain) fault of Montana (page 376), one of the best known overthrust faults in the United States, has a displacement of 10 miles or more.

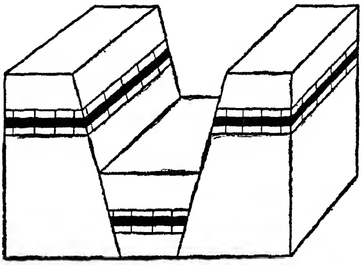

FIG. 401.—Diagram showing two normal faults with a depressed block or *graben* between them.

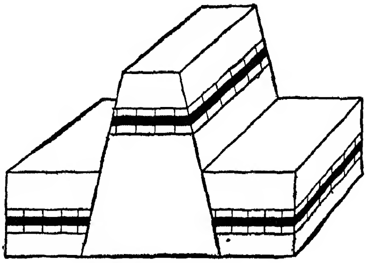

FIG. 402.—Diagram showing two normal faults with an elevated block or *horst* between them.

Topographic Expression of Faults.—A fault scarp is a steep slope or escarpment that has formed along a fault. When faulting takes place, the upthrown block generally is higher than the downthrown block. But erosion works faster on the elevated than on the depressed one, because it is higher. Thus the cliff due to faulting is worn back and recedes from the outcrop of the fault. A mountain block raised by faulting may be no higher at the fault line than the neighboring block that was relatively depressed.

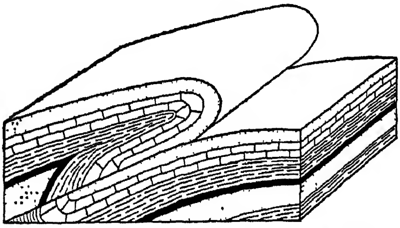

FIG. 403.—A recumbent fold passing into a thrust fault.

Unconformity.—Where rocks are laid down one above the other in uninterrupted succession, they are comformable. When they are eroded and submerged and later rocks are laid down upon them, the relationship is an unconformity (Fig. 404). The plane of contact between the two beds or series of beds is the plane of unconformity. Commonly this is not a geometrical plane but an undulating one. If the beds below and above the plane of unconformity have the same dip, the relation is an erosional unconformity or a disconformity. If the beds of the lower series were folded or tilted before beds of the later series were laid down, the two series of beds are discordant. The beds

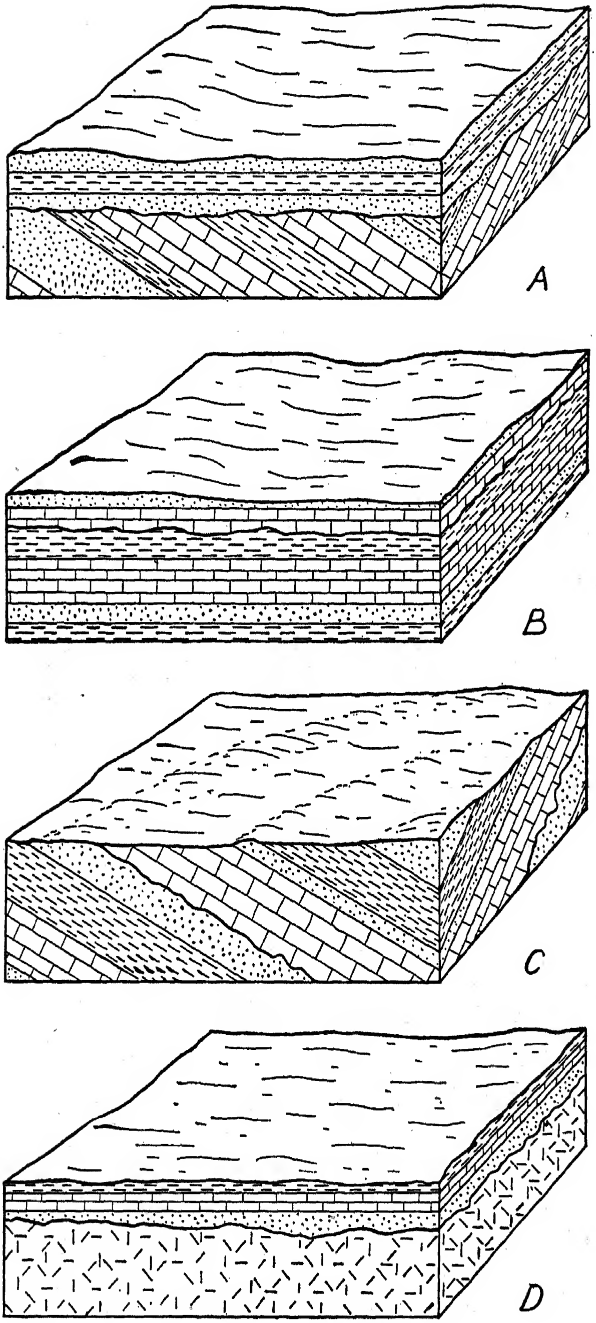

Fig. 404.—Diagrams illustrating unconformable relations of sedimentary rocks. *A*, an unconformity with a discordance in bedding; *B*, an erosional unconformity or a disconformity, neither series of strata is tilted or folded; *C*, a disconformity with both series tilted; *D*, sedimentary rocks unconformably above the eroded surface of igneous rocks.

dip at different angles, and the unconformity is an angular unconformity.[1] This type of unconformity is illustrated by Figs. 404*A* and 405.

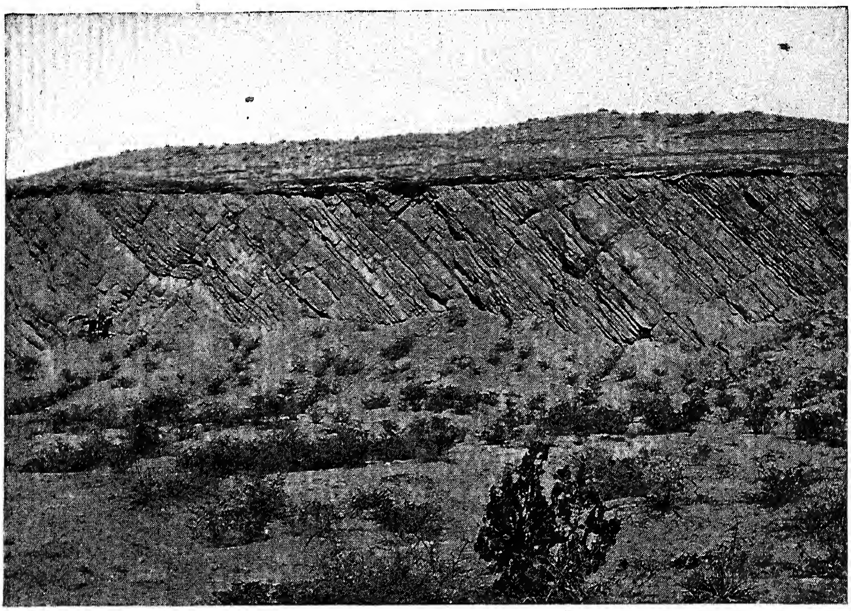

Fig. 405.—Unconformity near Socorro, New Mexico. (*Photograph by Chapman. U. S. Geol. Survey.*)

Unconformities may be developed between an eroded igneous or metamorphic rock and sedimentary rocks. An unconformity indicates a long chain of geologic events, and the relation therefore is highly significant (page 278).

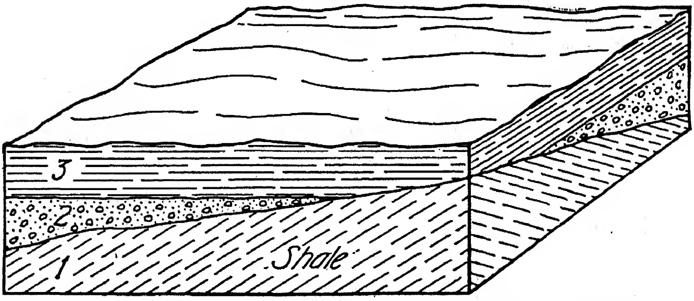

Fig. 406.—An unconformity with overlap. The oldest series consists of shales (1) which were folded, eroded, and partly submerged. A conglomerate (2) was deposited on parts of the eroded surface. Subsidence followed and shales (3) were deposited on the conglomerate (2) and also on the older shales (1).

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[1] Leith, C. K., Structural Geology, pp. 156–163, Henry Holt & Company, Inc., 1923.

When a bed or series of beds are laid down in a sea that is expanding, the later beds cover a larger area than the earlier ones. They overlap the earlier ones and cover surfaces that the earlier ones do not cover. This relationship is an unconformity by overlap. Figure 406 shows an area in which a bed (2) was laid down on a folded shale. Later a larger area was submerged, or the sea expanded and bed 3 was laid down on bed 2 and also on the older shale. Bed 3 is said to overlap bed 2.

Evidences of Unconformity.—The evidences of unconformity include all evidences of erosion between the deposition of the older and younger series of rocks. Thus, if the dividing surface between the older and the younger series is irregular (Fig. 404), it may suggest that it was once an erosion surface. An ancient soil or weathered zone between two series of rocks shows that there is a buried erosion surface. If the lower series of rocks is more highly folded than the later one, there is a discordance of bedding which indicates unconformity (Fig. 404*A*). If the lower series of rocks contains many veins or dikes and these are absent from the upper series, one may infer that an erosion surface exists between the two series. If sedimentary rocks lie upon a surface of igneous rock and there are no evidences of intrusive relations, it is probable that the contact is an erosion surface.

An abrupt change in the character of the rocks suggests that a change of conditions occurred. Thus a conglomerate, which is a near-shore sedimentary rock, suggests an ancient shore line. If the conglomerate contains fragments of the underlying formation, it is evidence of an erosion interval. A basal conglomerate at the beginning of a new series, containing fragments of different rocks from the older series, particularly is significant. Conglomerates are found within formations, and some of them seem to be made up of fragments of rocks broken by waves during heavy storms. These intraformational conglomerates do not denote unconformities and are to be distinguished from basal conglomerates.

SUMMARY OF EVIDENCES OF UNCONFORMITY

1. Discordance in bedding. The underlying series of beds is more highly folded than the overlying series.
2. Erosion surface. The beds of the upper series rest upon an erosion surface of the older series.
3. Basal conglomerate. The lowest beds of the upper series contain pebbles of rocks of the older series.
4. Differences in degree of deformation. The older rocks are faulted, folded, or metamorphosed more than the younger series.
5. Differences in veining and intrusion. The older rocks contain closely spaced dikes and veins not present in younger series.
6. Differences in character of rocks. An intrusive igneous rock is in contact with sedimentary beds but does not exhibit intrusive relations, showing that it was eroded before the upper beds were deposited.
7. Marked differences in fossils occurring in the beds in contact.

Causes of Diastrophism.—The orogenetic and the epeirogenetic movements are probably due to shrinkage. The interior of the earth is continually losing heat. This is certain, because measurements in wells and mines show that the earth gradually becomes hotter with depth, and heat moves from hot to cold regions. Rocks contract or shrink when they cool, and they contract also by forming denser materials. The rocks inside the earth become denser. The outer shell or crust of the earth is warmed by the sun; it remains essentially at a constant temperature; and moreover it is under less pressure. The shrinkage of the inner ball of the earth that lies within the shell or crust would tend to cause openings to form, but the rocks are not strong enough to hold the walls of openings permanently apart. Therefore, any large openings that were formed would be closed by the great pressure of overlying material. Since the crust does not shrink from loss of heat, it must fail or crush in order to fit the inner ball. By its failure the crust is warped, and mountain folds are formed when the crust, or outer shell, which is too large to fit the inner ball is readjusted to fit the smaller ball which it surrounds.

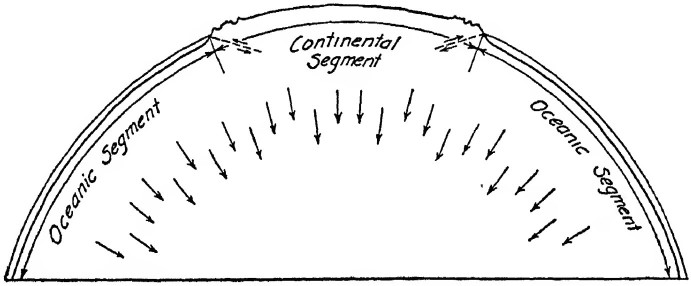

FIG. 407.—Cross section of part of the earth to illustrate its deformation near the borders of a continent. The arrows that point to the center of the earth indicate its condensation due to gravity. The long arrows (arcs) represent tangential stresses near the earth's surface. Each of the three segments is sinking, but the central continental segment is wedged up relative to the oceanic segments. The dash lines represent shear zones that pass under the edges of the continental segment. Radial lines represent the relations of segments before deformed. (*After T. C. Chamberlin, R. D. Salisbury, and R. T. Chamberlin.*)

A commonly accepted theory of earth deformation is that the oceanic and continental segments of the crust act as wedges. These wedges move downward (Fig. 407); but since the oceanic segments are greater and are composed of heavier material,[1] they move downward more than the continental wedges, whereas the latter are squeezed between the ocean segments. This causes thrust near the borders of the continental wedges, where the mountain folding generally is concentrated. By

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[1] CHAMBERLIN, T. C., and R. D. SALISBURY, Geology, vol. I, pp. 551–569, 1905. A clear treatment of the subject is found in College Geology, by the same authors, revised by R. T. Chamberlin and Paul McClintock, vol. I, p. 288, 1927.

shrinkage of the earth's interior, stresses gradually accumulate; when a point is reached so that the crust can no longer withstand the stresses, the crust yields.

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CHAPTER XIV

THE ORIGIN AND STRUCTURE OF MOUNTAINS

The word mountain is a term applied to all eminences that rise to considerable heights above their surroundings. There is no sharp distinction between hills and mountains. An elevation that appears to be no more than a small hill in the rugged portions of the Rocky Mountains would seem to be a great mountain if transported to the plains of the interior of the continent. Mountains differ in their modes of origin and in their life histories. One may be eruptive, resulting from the extrusion of lavas. Another may be the result of the intrusion of an igneous body into other formations, the latter being eroded away from the igneous mass after it has cooled and hardened, leaving it standing out as a hill or mountain. Still another may be the result of direct uplift above the surrounding region by lateral squeezing. Others still are eroded from broad, level plateaus raised above the surrounding regions sufficiently to provide a more rapid degradation. Mountains are formed also as the result of simple folding or tilting of blocks produced

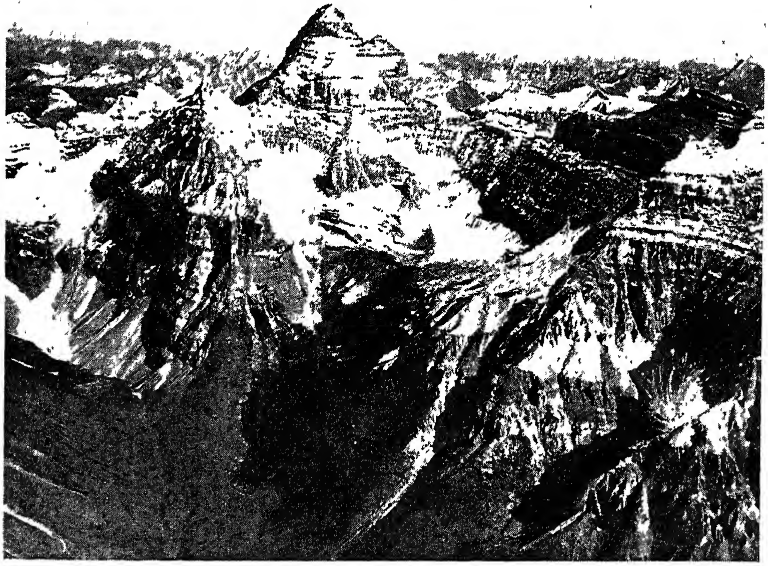

FIG. 408.—Mount Assiniboine, Alberta, Canada. A mountainous area carved from elevated flat-lying and gently tilted beds. (*Courtesy Royal Canadian Air Force.*)

by faults. Many mountains are formed by two or more of these processes working together (Fig. 408). Adding to these the variation in the character of the rocks, the different positions to which strata are tilted or folded, and the effects of these differences upon erosion, it is readily understood why mountain structures are so varied in different localities.

Modes of Occurrence.—Mountains occur as isolated peaks, irregular groups, parallel ridges or ranges, and complex systems. The broad, mountainous belt in western North America, from the eastern border of the Rocky Mountains to the Pacific Coast, is commonly referred to as the North American Cordillera. A similar belt in southern Europe and Asia is designated as the Eurasiatic chain or mountain zone.

Mountain peaks are high masses, more or less conical in outline, that rise above their surroundings. They may be either mountains of accumulation, such as the volcanic cones on the floor of the sea or along certain coasts of the Pacific Ocean, or erosional remnants, such as Mount Monadnock in New Hampshire or Stone Mountain in Georgia.

Irregular groups of mountains vary in size and arrangement from small mountainous areas like the Little Rocky Mountains of Montana or the La Sal Mountains of Utah to larger, irregular units such as the Front Range of Colorado and Bighorn Mountains of Wyoming.

Mountain ridges and ranges are long, narrow, mountain masses that may represent the arches of anticlinal folds or the outcropping ridges of resistant rocks, remaining after the folds have been deeply eroded. Such folds are seen in the Coast Range, the Wasatch Range, the Pyrenees, and other mountain ranges.

A mountain system consists of several more or less parallel ranges in the same region. Thus the term Laramide system often is used in referring to a series of the ranges of the Rocky Mountains.

Residual or Erosional Mountains.—Mountains may be carved from a featureless plateau. Where diastrophic movements raise broad plateaus above the surrounding region, the gradient of the streams is increased, and the rate of stream erosion is accelerated. In time the plateau becomes extensively dissected by steep-walled canyons, and the ridges and pyramids representing the remnants of the plateau stand so high above the stream valleys that they are mountains. An early stage of such mountain building is seen in the region of the Colorado plateau in Arizona, where gorges more than a mile deep have been cut by the Colorado River and its tributaries. The divides between tributaries are irregular, flat-topped mesas and sharp ridges that are notched or broken into a series of isolated peaks and truncated pyramids that rise from the depths of the gorges. Their bases are far below the plateau, and they are dwarfed by the magnitude of the plateau from which they are carved. At Zion Canyon, Utah, horizontal strata lying 4,000 to 8,000 feet above sea level are cut by canyons several thousand feet deep (Fig. 409). In this region many of the pyramids that represent the divides between smaller tributaries exhibit well-developed rock terraces separated by precipices and long, vertical cliffs.

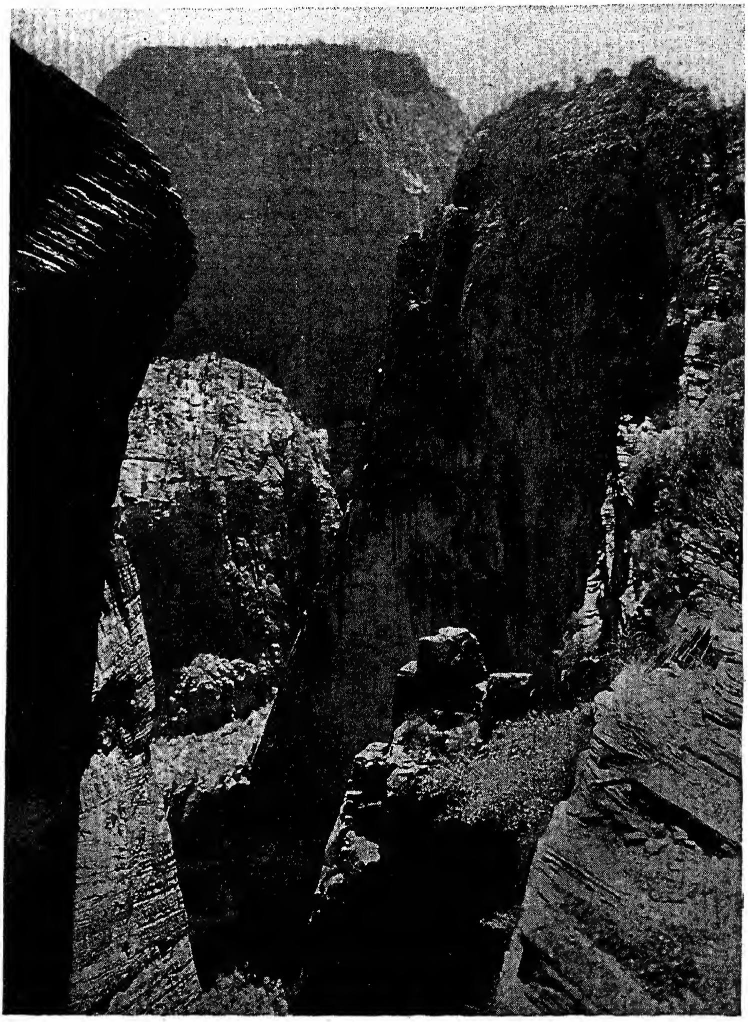

Fig. 409.—Residual mountains carved from the pyramids between tributary valleys. Zion National Park, Utah. (*Courtesy Union Pacific R. R.*)

Residual mountains may be formed also from plateaus of homogeneous crystalline rocks or of tilted, folded, or faulted strata that represent the elevated or rejuvenated surface of a former peneplane. The form of the mountains carved from such plateaus depends upon the nature and arrangement of the rocks out of which they are sculptured.

Volcanic Mountains.—Volcanoes may occur singly or in irregular groups (Fig. 410). They vary in size from low mounds to lofty peaks. Some are built up in low-lying regions, but commonly they are superimposed on the crests of mountain ranges that owe their origin to other processes. The form of the cone depends very largely upon the nature of the materials of which it is composed. Most volcanic mountains consist of successive layers of ash, dust, and coarser pyroclastic fragments, interbedded with sheets of lava. Those that eject great quantities of solid rock fragments, as a rule, have higher and steeper peaks than volcanoes which emit highly fluid lavas. Many of the highest peaks in the world, such as Aconcagua (23,080 feet), Chimborazo (20,498 feet), and Cotopaxi (19,613 feet) in the Andes and Kilimanjaro (19,710 feet) in Africa, have been built mainly by volcanic processes. In the United States typical examples are Mount Shasta (14,380 feet), Mount Rainier (14,408 feet), and Lassen Peak (10,437 feet).

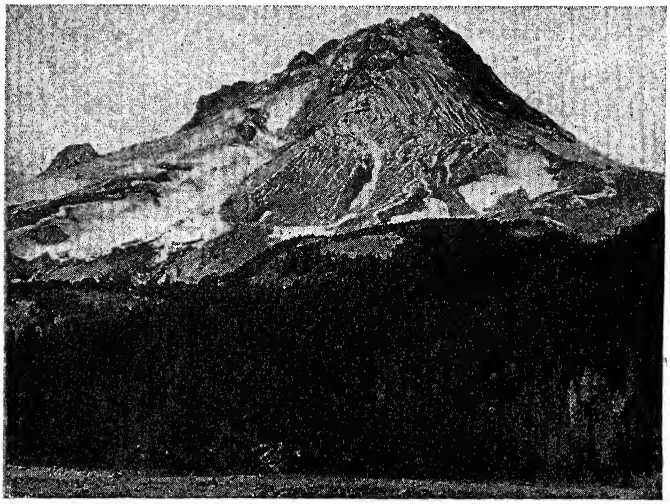

FIG. 410.—Mount Hood, Oregon. A volcanic mountain peak as seen from White River glacial moraine. (*Courtesy Union Pacific R. R.*)

Mount Shasta rises almost 2 miles above its base. It is a volcanic cone composed of alternating lavas and tuffs. The base of the cone at about 5,000 feet above sea level is 17 miles in diameter. It has an average slope of about 15 degrees. The slope of the upper third is about 35 degrees, but this flattens toward its base, where the slope is less than 5 degrees. The volume of the cone is estimated to be more than 80 cubic miles.

Volcanic action may become so general along definite lines or zones on the earth's surface that whole mountain ranges may result from the accumulation of volcanic débris. This is shown along the chain of the

Aleutian Islands of Alaska, where a mountain range over 1,000 miles long is now being built on the floor of the sea, mainly through volcanic activity. Vulcanism played a part also in the formation of the Cascade Range, extending from northern California northward into British Columbia. Along this zone Mounts Shasta, Lassen, Pitt, Hood, Adams, Saint Helens, Rainier, Baker, and other volcanoes contributed enormous quantities of volcanic materials during the formative stages of the mountain range.

Moraines and Dunes.—Glacial moraines and dune ridges represent the heaping up of materials at the surface by gradational agents. They rarely attain the dimensions implied by the term mountain. Many sand dunes are 300 feet high, and, under exceptional conditions, they reach heights of over 1,000 feet. Moraines commonly are 200 to 300 feet high but are rarely more than 1,000 feet above their bases. Although certain authorities regard moraines and dunes as mountains, this practice is not general.

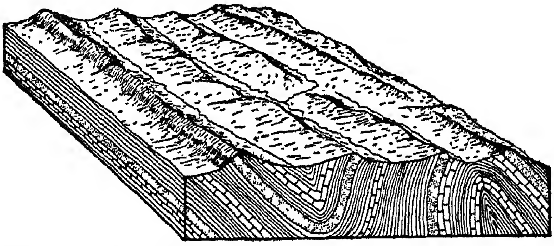

FIG. 411.—Erosional mountain ridges formed by the differential erosion of folded sedimentary strata. (*After Matthes, U. S. Geol. Survey.*)

Folded Mountains.—The greatest mountain systems of the earth were made by the folding of rock strata. The Himalayas, the Carpathians, the Alps, the Pyrenees, the Andes, and the Rocky Mountains are all folded mountains (Figs. 411, 412, 413). Folded mountains show various types of folds, such as the simple dome-like anticlines of the Uinta Mountains of Utah, gently folded strata such as characterize the Jura Mountains of Switzerland, or the highly compressed and crumpled fan folds of the Alps.

The Uinta Mountains in northeastern Utah represent the simplest type of folded mountains. This range is approximately 150 miles long and 40 miles wide, with many peaks and crests more than 12,000 feet high. The structure is a flat-topped anticline between abrupt monoclines. Dips in the central platform average about 5 degrees, while in the flanking monoclines the strata are inclined about 45 degrees. A thickness of over 3 miles of sedimentary rocks has been eroded from the top of the broad anticlinal arch, and the numerous peaks that remain are sculptured from resistant Pre-Cambrian quartzite.

The Jura Mountains lying along the border of France and Switzerland are symmetrical open folds. Here the folds are eroded very little, and the ranges consist of a series of anticlines, and the valleys are synclinal troughs. Similar folds are observed where the Appalachian Mountains cross Pennsylvania, Virginia, and Maryland (Fig. 411). However, these folds are deeply truncated by erosion, and consequently the ranges are long, parallel, sharp-crested ridges with narrow valleys intervening. The ridges are developed upon the flanks of anticlines, where resistant strata are exposed, or along the axes of synclines, where stream adjustment has shifted the valleys to less resistant strata (Fig. 411). Where the axes of the folds are not horizontal but pitch or plunge into the earth, a zigzag pattern of ridges is developed. Outstanding examples occur in Pennsylvania at the western end of the anthracite coal region, where the ridge crests zigzag back and forth in a series of loops.

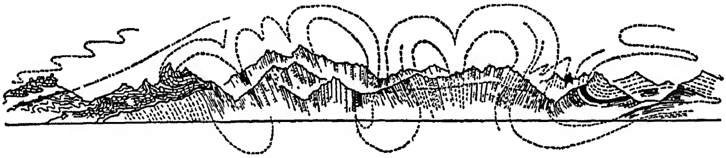

FIG. 412.—Cross section of the St. Gotthard massif, Switzerland, showing the fan folds that are characteristic of the Alps. (*After Heim.*)

In mountains where compressional forces were more active the folds are closely compressed and are inclined at all angles. In a great mountain chain like the Alps such crumpling occurs on a gigantic scale. Intense lateral thrusts have overthrown enormous folds that have thrust old crystalline rocks high above younger sedimentary strata. In several districts distinct loops, rising one above the other, can be seen in the great cliffs which border the valleys. Some of the highest summits, such as Mont Blanc and the Jungfrau, have been carved from similar structures. In the central portion of the Alps many of the folds that were arched upward were squeezed and compressed to such an extent that they became overturned on either side of the central axis, producing huge fan-shaped folds (Fig. 412).

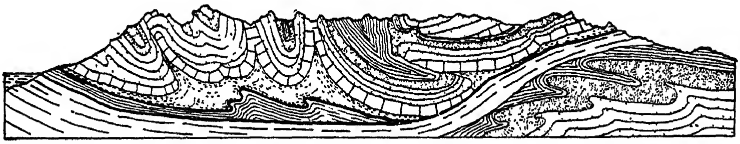

FIG. 413.—Cross section through the west Alps showing the complexly folded and faulted strata that have been thrust over younger beds. (*After Schardt, from Kayser's Lehrbuch der Geologie.*)

Faulted Mountains.—Complex folding usually is accompanied by fracturing. Where the rocks are broken by continuous fractures,

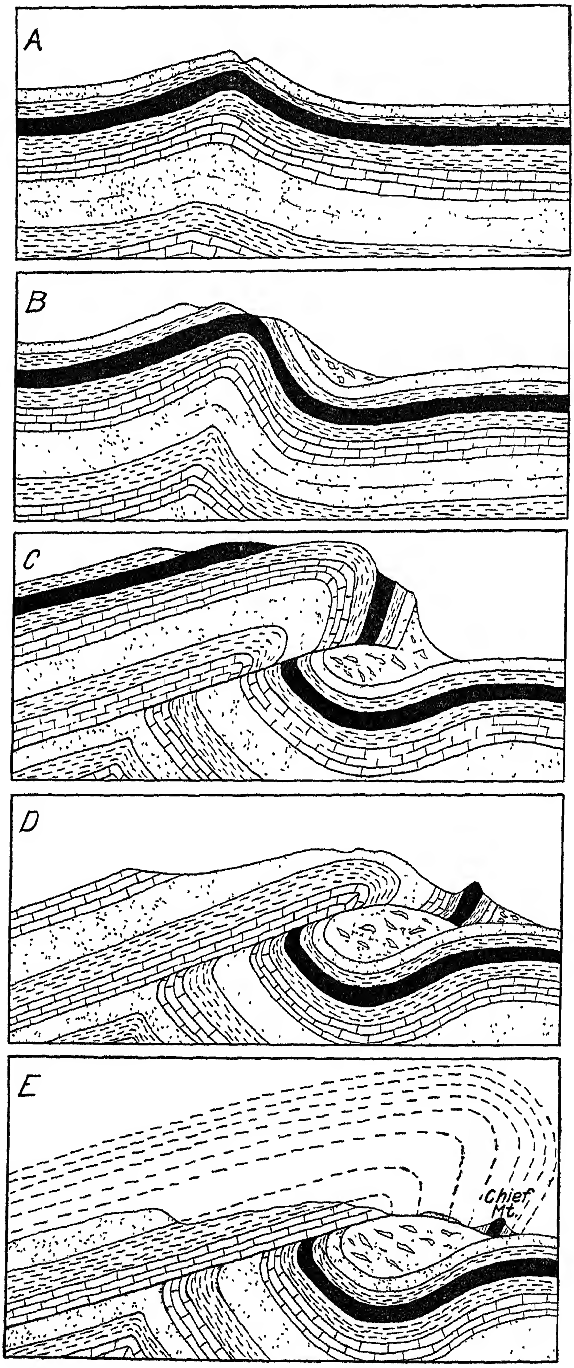

FIG. 414.—A series of diagrams showing the probable sequence of events (*A* to *E*) in the formation of the Lewis overthrust in northwestern Montana. Old geologic formations were thrust eastward more than 10 miles along a low-angle fault plane overriding younger strata. Chief Mountain is an erosional remnant of the ancient Pre-Cambrian

displacement or faulting may take place as deformation continues. Most intensely folded mountain structures show that both folding and faulting were involved in their formation (Fig. 413). Thrust faulting is a very common structural feature in mountains formed by compressional forces. The Highlands of Scotland are illustrations of mountains built by a shortening of the earth's crust through a series of distributive overthrust faults. In the United States both the southern Appalachians and the northern Rockies show overthrust faulting on a large scale. In northern Montana the Front Range is represented by the Lewis and the Clark ranges. Both ranges have high, craggy peaks reaching altitudes from 6,000 to 10,000 feet. The whole structure is a synclinal block, each limb cut off by a steep outer face. The two ranges are formed by the upturned edges of the strata, dipping from 5 to 30 degrees toward the central trough. The entire block of the earth constituting the ranges has been thrust eastward by one of the greatest faults known. Ancient dolomites, quartzites, and argillites which make up the mountains have been pushed eastward over the younger Cretaceous shales of the plains. The thrust plane is nearly horizontal. It can be traced 7 miles in the direction of thrusting, but the extent of the thrust was greater, because the overthrust block has been reduced by erosion. Field evidence shows that the thrust was at least 10 miles horizontally, and it may have been much greater[1] (Fig. 414).

Faulting and crustal warping of great magnitude resulted in the formation of the Sierra Nevada on the eastern border of California. These mountains form a continuous range about 75 miles wide and nearly 400 miles long. They represent a huge block of resistant rock uplifted by faulting and tilted toward the west. For this reason the crest of the range is near its eastern margin, and the eastern slope is steep and cliff-like. Where the crest line is highest, the average eastward slope from crest to foot is more than 1,000 feet per mile. In contrast to the steep eastern front is the gentle westward slope, that descends gradually from 9,000 feet on the east to 1,000 feet above sea level at the margin of the central valley of California. Along the crest of the range Mount Whitney rises to 14,495 feet, the highest peak in the United States outside Alaska. All structural features of the east front of the range indicate that it is a huge fault scarp. The displacement did not take place along a single fracture but rather along a zone with numerous compound faults that have left spurs and offsets in the present topography.

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[1] WILLIS, BAILEY, Stratigraphy and Structure of the Lewis and Livingston Ranges, Montana, *Bull.*, Geol. Soc. America, vol. 13, p. 307, 1902.

FENNEMAN, N. M., Physiography of Western United States, p. 206, McGraw-Hill Book Company, Inc., 1931.

The Vosges and Black Forest mountains that form the walls of the Rhine valley are fault-scarp mountains (Fig. 415). These mountains have escarpments that face each other across the valley. Before the valley was formed, the two ranges probably were welded together as a broad anticlinal ridge, composed of a core of ancient crystalline rocks overlain by younger sedimentary strata. Faulting caused the crest of the arch to subside, forming a trough-like valley by the depression of a series of small fault blocks.

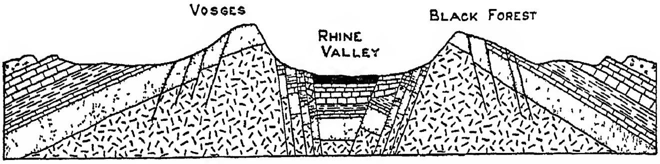

Fig. 415.—Cross section of the Vosges and Black Forest ranges that parallel the valley of the Rhine. Vertical scale greatly exaggerated. (*After Penck.*)

Block Mountains.—Where faulting takes place on a large scale, mountains may be produced by differential movement along fractures in the earth's crust. This may be accomplished by elevation or by depression along one side of a fracture, leaving the other side in its original position, or by a combination of the two types of movement. Where there are great systems of intersecting or parallel fractures along which there has been differential movement, mountains known as fault, or block, mountains are produced (Fig. 416). In most cases, block mountains represent segments of the earth which probably have maintained their positions while the adjoining blocks have broken away and subsided. The most notable examples of this type of mountain structure are found in the Great Basin area in the southwestern part of the United States. This region is bordered on the east by the Wasatch Range and on the west by the lofty Sierra Nevada. It extends for nearly 800 miles north and south and for 500 miles east and west. The basin area is traversed by numerous approximately parallel mountain ranges, with their axes somewhat irregular but generally extending north and south.

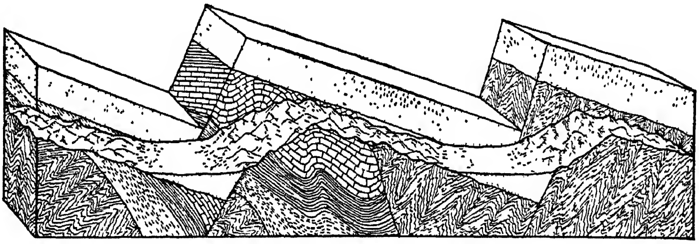

Fig. 416.—Cross section of erosional mountains sculptured from block-faulted strata. The background shows diagrammatic fault escarpments prior to erosion. In the foreground the same blocks are shown eroded to form mountains. (*After W. M. Davis.*)

Many of the ranges rise from 3,000 to 5,000 feet above their bases and are outlined by normal downthrow faults the escarpments of which are but slightly altered by erosion. The abrupt change from valley floor to mountain slope together with the uniform slopes of the mountain sides is a striking characteristic of basin and range structure. In the more arid portions of the basin the slopes range from 20 to 90 degrees. In the northern part of the basin many of the fault blocks have been so deeply dissected by erosion that their original form has been almost completely destroyed. In this region many of the intermontane depressions are so filled with rock waste deposited by intermittent streams that some of the mountains are partly buried by alluvial cones and talus slopes of débris derived from the weathering of the fault escarpments.

Domed Mountains.—In a number of mountainous regions where the mountains occur as irregular groups, the central area of the group is composed of intrusive rock, such as granite, around which sedimentary rocks are found dipping in all directions away from the igneous core. The structure of the strata indicates that they once formed a continuous roof over the igneous mass. Such structures are formed where magma is injected into the earth's crust in such quantity that the surface strata are lifted into dome-like structures. Erosion removes the strata from the top of the dome, and irregular hills and mountains are carved from the exposed igneous core. In other mountain groups ancient igneous bodies such as granite masses were laid bare by erosion and subsequently covered by sedimentary beds.

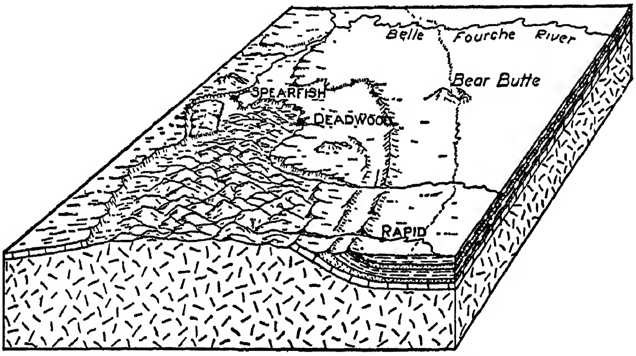

FIG. 417.—Block diagram of northern part of the Black Hills, South Dakota. A group of mountains formed by the dissection of a dome-like uplift. (*In part from Lobeck.*)

The Black Hills in southwestern South Dakota and the adjoining portion of Wyoming are mountains which rise several thousand feet above the level of the surrounding plain (Fig. 417). They are carved from a dome-like uplift that is nearly 100 miles long and approximately 50 miles wide. Before the sedimentary rocks were removed from the top of the dome, it must have risen at least 6,000 feet above the plains.

The exposed ancient core of the dome is composed of granite that has been eroded to form many ridges and peaks that culminate in Harney Peak, which rises to more than 7,000 feet above sea level. Around the base of the igneous core is a ridge of limestone younger than the granite with a steep escarpment facing inward toward the center of the hills. Beyond the limestone outcrops are hogback ridges of sandstone, which constitute the outer rim of the dome-like structure.

The Henry Mountains in southern Utah are typically laccolithic structures. They stand singly or in clusters upon a desert plain that has an altitude of over 5,000 feet. They are a group of five individual mountains the highest of which is Mount Ellen, which reaches over 11,000 feet above sea level. The laccolithic domes show considerable diversity in the amount of erosion. Some are still partially or completely capped by overlying sedimentary strata, whereas in others the intruded igneous rock is exposed on the crest of the structure and the sedimentary strata appear in hogbacks or cuestas that encircle the uplift.

History of Mountain Systems.—The history of certain folded mountain systems may be divided into three definite stages, each of which shows the operation of distinctly different processes: (1) the preparatory stage during which thick sediments accumulate in slowly subsiding geosynclines; (2) the period of deformation during which the strata are subjected to lateral pressure and are folded into mountains; (3) the period of denudation accompanied or followed by vertical earth movements that tend to revive the mountains. These stages are discussed in paragraphs that follow.

The sedimentary rocks included in folded mountains commonly contain a high percentage of coarse clastic sediments, such as sandstones and conglomerates. Certain beds or groups of strata traced for some distance from the mountains show both a decrease in thickness and a gradation in texture to finer materials. In the Appalachian Mountains the Paleozoic sediments have a thickness of about 25,000 feet, whereas in the Mississippi Valley the same series of rocks are approximately 5,000 feet thick. The nature of the rocks included in the mountain folds indicates that they are of shallow-water origin and that they were deposited under near-shore conditions of sedimentation. Since the sediments have a total thickness of about 25,000 feet, the basin in which they were accumulating must have subsided while deposition was in progress. Structural relations in the series indicate that sedimentation was interrupted by uplift and erosion. Such periods of uplift rejuvenated the marginal land areas and thereby increased the supply of sediments. These processes continued until the subsiding trough became an elongated geosyncline extending from Alabama northward into New England. Most of the sediments brought to this trough-like basin came from an ancient land mass that was to the east of the geosyncline and extended eastward to include the area of the present continental shelf (see Fig. 418). This land area continued to rise as the geosyncline was depressed, and eventually erosion brought to the surface and exposed crystalline rocks that were massive and resistant. Great remnants of these are exposed in the piedmont area just east of the Appalachian Mountains.

The second stage in the development of the Appalachian Mountains began when the weaker sedimentary strata in the depressed trough began to yield to lateral pressure tangential to the earth's surface (see pages 339 to 369). The diastrophic forces caused the sediments to be pushed together against the resistant land mass, buckling the beds and arching upward huge anticlinal folds. Where the compressional forces were most intense, close isoclinal folds were formed from broad anticlinal arches the crests of which were pushed upward as the limbs of the folds became more nearly parallel. Such upward movement made room for more lateral movement and resulted in a great shortening of the earth's crust (see Fig. 382). During the period of active deformation the moun-

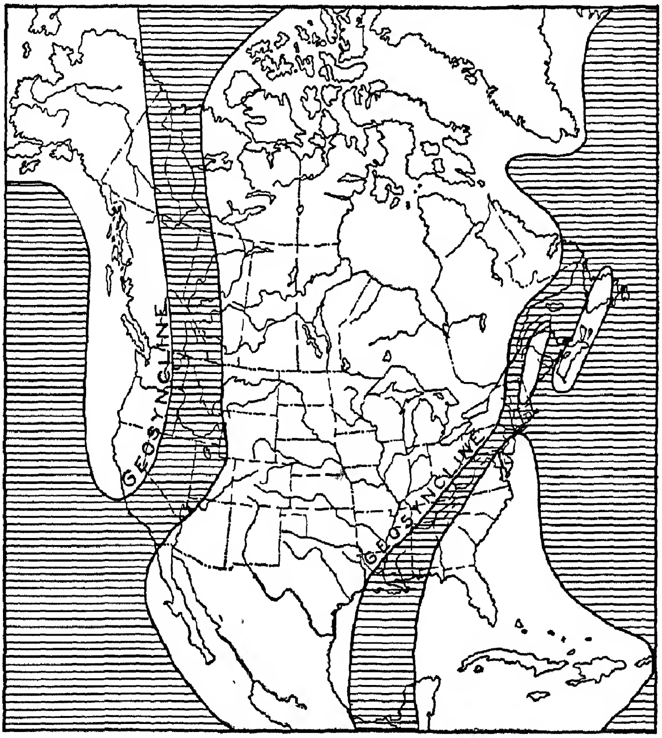

FIG. 418.—Map of North America showing the location of early Paleozoic geosynclines in which the sediments accumulated that were later converted into mountains by diastrophic processes. (White areas, land; shaded areas, sea.)

tains continued to rise. Some parts were raised higher than others by folding or faulting, and some adjoining areas were lowered. Sudden displacements may have caused local shifting of stresses within the earth so that conditions were favorable for the rise of molten rock and volcanic forces became active. Since the rate of uplift was slow and the rate of denudation was increased by the elevation, erosion became a factor in sculpturing the mountains.

The amount of compression represented by a folded mountain system can be estimated by detailed field investigations. The strata composing the northern Appalachian Mountains were compressed approximately 100 miles. In other words, the earth's superficial crust was shortened about 100 miles. If the folded strata of the Alps could be smoothed out into horizontal beds, they would cover an area about 125 miles wider than the zone occupied by the present mountains. In the zone of greatest compression, the reduction has been one-half or more. In some folded mountains much of this reduction is accomplished where overturned folds are broken and reverse or thrust faulting takes place. Such structures are common in the Front Range in Montana and Alberta and in the southern Appalachians.

The third stage in the history of a folded mountain system begins when deformation ceases or when the rate of denudation exceeds that of uplift. During this stage the degradational work of erosion reduces the height and ruggedness of the ranges, and finally the mountains are lowered by erosion to a peneplane. The resulting peneplane, however, rarely represents the final extinction of a mountain system. Subsequent uplift may rejuvenate the streams, and mountainous features again are etched out of the old topography. The forms developed in the revived mountains are determined by the amount of uplift and the nature of the rocks. In the Appalachian Mountains the peneplaned stage was followed by uplift, accompanied by warping and arching. Thus some areas were elevated higher than others. This is illustrated by the Piedmont plateau, which is low and uniform, while the Blue Ridge range along its west border was lifted to mountainous heights. Still farther west in the younger Appalachian folds the former peneplaned condition is very evident. There even-crested ridges that represent the surface of the former peneplane have been etched into relief and extend for miles with remarkably uniform elevation. The ridges are separated by adjusted stream valleys that parallel the strike of the weaker strata. Some of the streams inherited meandering courses that have become intrenched as the region was rejuvenated.

Extinct mountains now reduced to peneplanes are found on all continents. They occur in areas of ancient crystalline rocks that have been deeply eroded because of successive periods of uplift. On the

North American continent the Laurentian peneplane contains a number of outstanding examples. In the Lake Superior region the stumps of the peneplaned Algoman mountains may be traced from north central Minnesota northeastward into central Ontario. Other ancient mountains have been submerged and covered with younger sediments and later have been reelevated as plateaus of horizontal strata resting upon the irregularities of the old erosion surface. Thus mountains come and go. They rise from synclinal troughs; their summits are pushed up to high elevations and later are eroded; and at places their stumps are buried under later formations.

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CHAPTER XV

METAMORPHISM

Agents.—Metamorphism includes all of those processes by which rocks are changed, whether chemically or by recrystallization and rearrangement of their mineral constituents. As practically all rocks have been changed to some extent since they were formed, the term metamorphism by many is limited to those more extensive changes involving the formation of new crystals or of new textures. Rocks are brought into new environments by processes of gradation and diastrophism. At places they are becoming more deeply buried; at other places they are being exhumed by erosion. The rocks and minerals that are stable under one set of conditions are unstable under another set, and they are changed to meet the new conditions. These and other changes are included in metamorphism.

The agents of metamorphism are chiefly water solutions, gases, pressure, and heat. On the basis of the agency most active in producing the changes the metamorphic processes may be classified as follows:

1. *Hydrometamorphism*, which includes the changes accomplished mainly by water at moderate temperatures and pressures.
2. *Hydrothermal metamorphism*, in which the changes are accomplished by vapors and by hot waters.
3. *Contact metamorphism*, in which the changes take place where an igneous mass invades an older rock and alters it at or near the contact.
4. *Dynamic metamorphism*, in which movement due to pressure is prominent, and rearrangements of constituents of the rocks are brought about through readjustments made by prolonged pressure.
5. *Static metamorphism*, in which the rocks, deeply buried, are under pressure of the overlying load of material, but movement due to pressure plays a minor part.
6. *Regional metamorphism* is a term applied where the rocks within a great area suffer profound changes. In this sense dynamic and static metamorphism also are generally regional.

Hydrometamorphism.—Some water is present in nearly all rocks. Near the surface, ground water brings new material in solution to the rocks; and by reactions of the water with the rocks, additions and subtractions continually are going on. Many of these changes are accomplished by *replacement*. In the process of replacement, solution f material is followed so closely by deposition that the *form* of the issolved body is preserved. An iron nail buried in soil will within a w years rust throughout (Fig. 419). It changes to iron oxide; and lthough it still retains the shape of the nail, it is no longer an iron nail; is so brittle and weak that it may e broken with the fingers. A trunk f a tree is replaced by silica so that e replacement or petrifaction shows ll of the textures of the trunk inuding bark, cells, etc. (Fig. 78). A rystal of iron sulphide is replaced by on oxide and yet retains the shape of e sulphide crystal. These replaceents are *pseudomorphs* (false forms), nd by studying them it is possible often to trace the changes rough which a mineral or rock has passed. At places fossils of lcium carbonate are present in a certain limestone, and in the etamorphosed equivalent of the limestone the fossils have been changed garnet, which retains the forms f the fossils. In some regions it is ossible to trace the altered sedientary rock into the unaltered sedientary rock and to follow the hanges step by step, for the grain, edding, and other textures of the original rock are preserved in the ltered rock after metamorphism. Figure 420*A* is a fossil shell. Figure 20*B* is a similar shell replaced by silver ore.

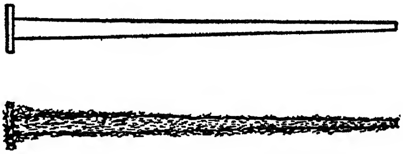


Fig. 419.—Drawing illustrating replacement. Upper figure shows an iron nail. Below is the same nail after oxidation. The oxidized nail has nearly the same shape as the unoxidized nail, but is no longer iron. It is so lacking in strength that it may be broken between the fingers.

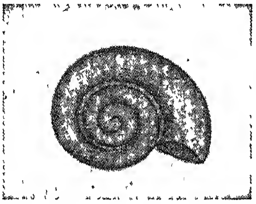


Fig. 420*A*.—Drawing of a fossil shell.

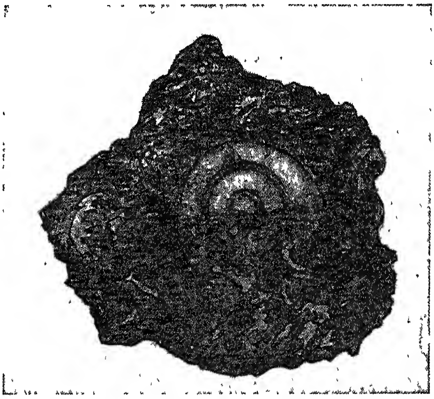


Fig. 420*B*.—Drawing of a fossil shell that has been converted by mineral bearing waters to silver ore. (*After Spurr.*)

Igneous rocks also are profoundly changed by replacement. Much aterial is removed, and other material takes its place. Commonly, owever, the texture or pattern of the rock is preserved, because the utlines of the original crystals form the outlines of the replacing crystals r of groups of the replacing crystals. By comparing the textures of the esh and altered rocks often it becomes evident that they were once like and parts of the same body.

Hydrometamorphism is an important process in rock weathering, but its effects extend much deeper than those of ordinary weathering. Surface waters carry downward into the earth carbon dioxide, oxygen, and other substances. The ferrous iron of igneous rock minerals combines with the oxygen and water, and a large part of the iron is converted into limonite. The alkalies and alkaline earths form soluble salts and are leached out of the rocks. Water is taken up in the reactions, and new hydrous silicates such as occur in clay are formed. These changes involve the formation of substances of lighter molecular weight than the original mineral matter. Consequently, the new minerals that are formed are, for the most part, of simpler molecular composition and of lower specific gravity.[1]

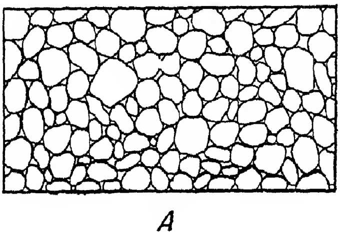


Fig. 421*A*.—Section of a sandstone poorly cemented or cemented by weak material such as calcite or iron oxide.

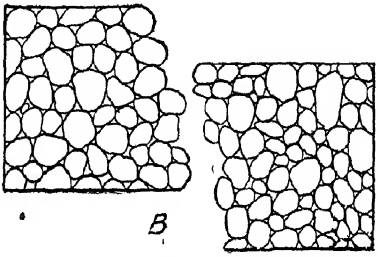
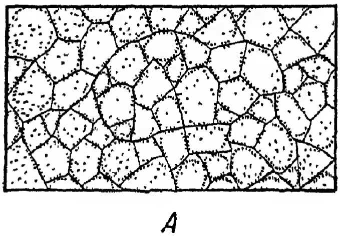


Fig. 421*B*.—The sandstone breaks due to fracturing of the cement. The quartz-sand grains are not broken.

Hydrometamorphism is active also in cementation. Underground water descending into porous rock carries mineral matter in solution. It consolidates calcareous sediments by forming minute crystals of calcite in them. A sandstone composed of quartz grains may be cemented by silica. Commonly the silica is added to the rounded particles of quartz so that each fragment of the earlier material is built up of similar material, and it would be impossible to ascertain where the original sand grain ends and later silica begins if it were not for minute dust



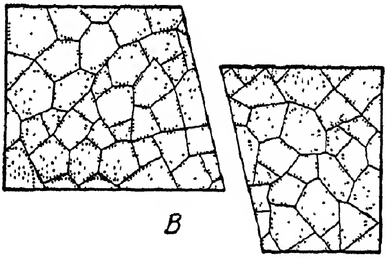


Fig. 422*A*.—Section of a sandstone cemented by silica. On deposition the silica enlarges the original quartz grains, eventually forming a solid body of silica. The sandstone by cementation thus becomes quartzite.

Fig. 422*B*.—Shows the quartzite broken across the grains.

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[1] The clay minerals are not of simple composition.

particles which separate the older and younger silica. Figure 421 shows an ordinary sandstone cemented; if fractured, it breaks through the cement and not through grains. Figure 422 shows a quartzite with dots of dust between old and new quartz; if fractured, the quartzite breaks across grains of sand and across the cement, since both are of the same material and are equally strong.

Hydrothermal Metamorphism.—Hot magmatic waters and associated gases are active agents of metamorphism. Magmatic heat makes ground water of meteoric origin more active. The high temperature and the high pressure under which magmatic solutions come out of intrusive magmas account for the pervasive character of this type of metamorphism.

Shales are converted to flinty hornstones; limestones are recrystallized, forming marbles; sandstones are changed to quartzites more commonly than by simple hydrometamorphism. Igneous rocks which are invaded generally are changed, and some are changed so much that they are recognized with difficulty. Glasses are devitrified or become crystalline and appear white and chalk-like. Granites are recrystallized, and additional mica is developed in them. These changes are accomplished by hot waters, i.e. by hydrothermal metamorphism. In such metamorphism replacement is common. Often such changes are attended by the deposition of ores of the metals. A well-known example of hydrothermal metamorphism is seen at Butte, Montana, a city built on granite. The waters that deposited copper ores altered the granite so much that practically no unaltered fragment is found in the vicinity of the ores. At many places in the earth's crust there are great bodies of rock composed chiefly of serpentine, a green mineral which is a hydrous silicate of magnesium. Many of these serpentines have formed from very basic igneous rocks, chiefly from peridotites which also are rich in magnesium. Hot waters have altered the magnesium minerals of the peridotites, chiefly olivine, which contains no water, and have changed them to the mineral serpentine, which contains water.[1]

Contact Metamorphism.—Where lavas flow over the surface, they may alter the rock over which they move. This alteration consists of melting or baking a thin layer of the older rock, but such changes are not profound in that they extend only a short distance, perhaps a few inches, into the older rock. The lavas contain gases, but these escape mainly upward. They may deposit minerals in vesicles or bubble

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[1] Magnesite is commonly formed along with serpentine. The formula is often written

Olivine + carbon dioxide + water = serpentine + magnesite

$$2Mg_2SiO_4 + CO_2 + 2H_2O = H_4Mg_3Si_2O_9 + MgCO_3$$

holes of lavas, forming amygdaloids, and, commonly, they cause some alteration of the rocks.

In intrusive rocks, however, the fluids are held in and, penetrating the walls of the intrusive chamber, accomplish greater changes. Along small dikes and sills the invaded rocks commonly are changed, but such changes, as a rule, are not very great, and they do not extend far away from the igneous body. Extensive changes are accomplished, however, by the gases which are given off by batholiths (Fig. 423), particularly in the invaded rocks near the stock-like masses of the upper parts of the batholiths. All rocks are changed—limestones, shales, sandstones, quartzites, and older igneous rocks, if such are present. The most profound changes are generally in limestone[1] and in calcareous shales. Calcium carbonate is changed to wollastonite by the addition of silica, or calcium carbonate is changed to garnet (andradite) by the addition of silica and iron, which are carried in the fluids that are expelled from the cooling magma. Shales are altered to hornfels. These changes are accomplished by replacement.

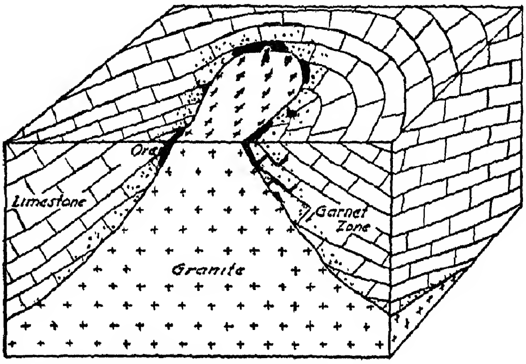

Fig. 423.—Diagram showing contact-metamorphic zone of garnet rock and sulphide ore deposits in the garnet zone. Such ores are confined to the invaded areas near the upper parts of the intrusive.

Pressure and dynamic movement of the invaded rocks, such as are attended by the formation of schists, play little part in contact metamorphism, and the products of contact metamorphism are rarely schistose unless the altered rock was a schist before contact metamorphism took place. The alterations are not caused by readjustments due to pressure

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[1] Limestone + silica = wollastonite + carbon dioxide
$CaCO_3$ SiO_2 $CaSiO_3$ CO_2
Dolomite + silica = pyroxene + carbon dioxide
$CaCO_3.MgCO_3$ $2SiO_2$ $CaMg(SiO_3)_2$ $2CO_2$
Limestone + silica + iron + water = garnet + carbon dioxide + hydrochloric acid
$$3CaCO_3 + 3SiO_2 + 2FeCl_3 + 3H_2O = Ca_3Fe_2(SiO_4)_3 + 3CO_2 + 6HCl$$

but by heat and mainly by hot fluids from the intruding rock. The contact zones around the intrusive at many places are 100 feet wide or more and at other places a mile wide or more. At many places, however, the altered zone is wanting.

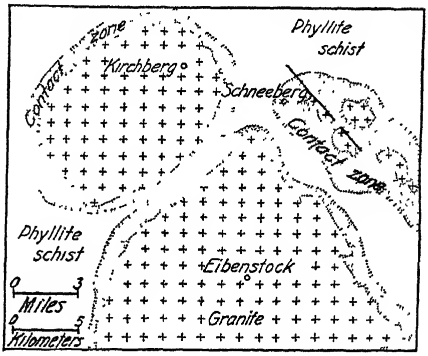

Fig. 424.—Map showing the granite masses of Kirchberg and Eibenstock, and the smaller stocks near Schneeberg in the Saxon Erzgebirge, Germany. The granite stocks are intruded into phyllite schist, which is metamorphosed to form "hornfels" or hornblende schist and other rocks. The smaller intrusives have the wider contact zones. (*After H. Müller.*)

Whatever material is present in the invaded rock generally is utilized in making up its metamorphosed equivalent. If the limestone contains magnesium, a magnesian mineral like pyroxene is formed; if clay or some other aluminous material is present in the limestone, the new minerals, such as pyroxenes and garnet, will be of the aluminous varieties. In most contact zones, however, much material is added. Silica and iron are very commonly introduced.

At many places where several intrusions are exposed in the same area, it is noted that the contact zone consisting of garnet rock or of hornfels is wider around the small, stock-like intrusives than it is around the larger ones. This is shown by Fig. 424. The small bodies near Schneeberg, in the Erzgebirge of Saxony, Germany, have wider contact zones than the larger Eibenstock mass, and the altered zone around the

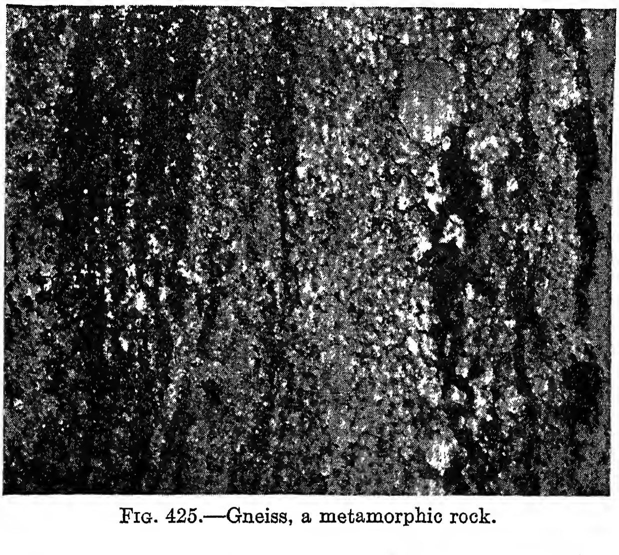

Fig. 425.—Gneiss, a metamorphic rock.

Kirchberg mass is of intermediate width. These outcrops represent bodies that probably increase in size with depth. They are the tops of cupolas, where the agents of metamorphism have been more active than lower down on the batholith.

Dynamic Metamorphism.—When rocks are deeply buried and are subjected to pressure and heat during great movements, they are in part recrystallized. When the term metamorphism is used without qualification, generally it refers to such recrystallization as has been induced by pressure and movement—that is, to dynamic metamorphism. All dynamic metamorphism implies movement. Rocks move differently in different parts of the lithosphere. The earth's outer shell may be

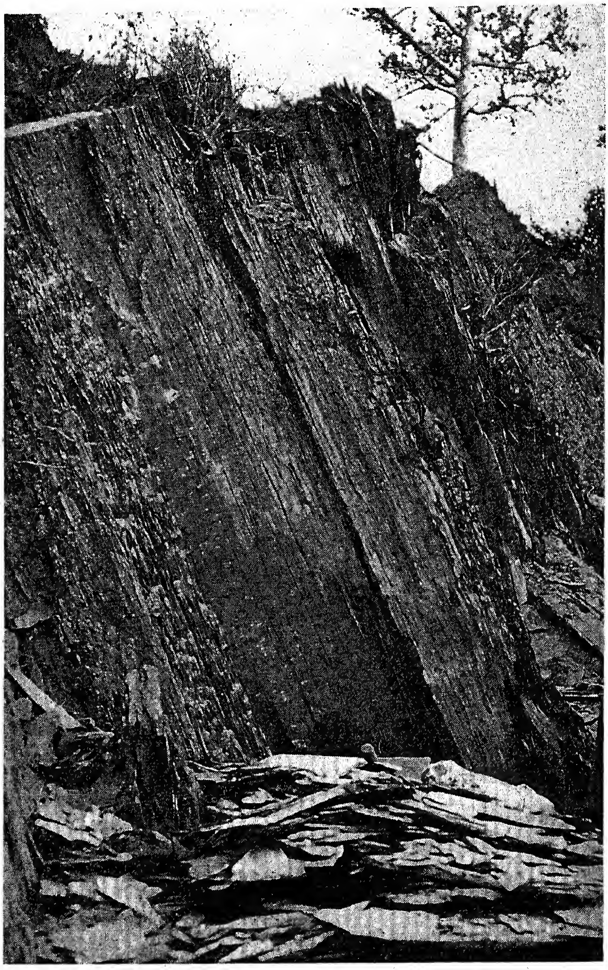

FIG. 426.—Slate, a mud metamorphosed by pressure. Deep Creek Canyon, 16 miles southwest of Townsend, Montana. (*Photograph by C. D. Walcott, U. S. Geol. Survey.*)

regarded as divided in two zones: (1) an outer zone near the surface, called the zone of fracture, where consolidated rocks will break; and (2) a deeper zone, called the zone of flowage, where even the stronger rocks are not strong enough to hold spaces open under the pressures that prevail.

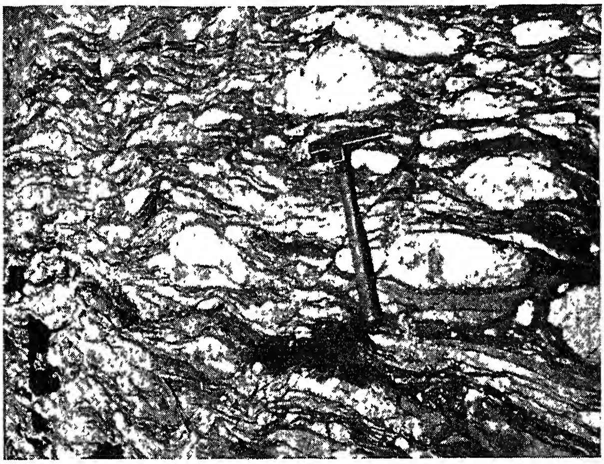

Fig. 427.—Conglomerate schist with pebbles elongated by rock flowage. (*Photograph by F. J. Pettijohn.*)

Rock flowage commonly attends dynamic metamorphism (see Figs. 425, 426, and 427). When the pressure on a rock is so great that small fractures, if formed, almost instantly are closed by pressure, the rock under deformation is said to "flow." The flowage of rocks under these conditions is not like that of a liquid. The rocks do not become even malleable like some metals. The movement takes place very slowly by the formation of minute fractures and gliding planes, by shearing, by granulation and recrystallization. The flow of glacial ice has been studied in great detail, as it illustrates probably the processes of flowage in other rocks. As with the movement of glacial ice, the factors causing flowage must operate chiefly below the melting temperature, and the rock mass remains essentially crystalline during the entire process.

Fig. 428.—Diagrams illustrating the deformation of marble by great pressure. The left figure shows a section of a hollow cylinder in which a small cylinder of marble is fitted. The right figure shows the same after great pressure has been applied by movements of the pistons. The deformation took place slowly, resulting in shortening and bulging out of the marble cylinder. It is deformed but it is not crushed. (*After Adams and Bancroft.*)

Rocks differ greatly in their response to pressure. Wet muds do not fracture but flow at the very surface of the earth, and soft shales will flow at shallow depths. Quartzites and igneous rocks are strong enough to hold fractures open even when there is a weight of several miles of rock above them. At greater depths, even stronger rocks yield by flowage. Experiments were made by Adams and Bancroft[1] in which they placed a cylinder of marble in a hollow cylinder of steel and after the application of great pressure on the pistons, as indicated by the arrows in Fig. 428*A*, *B*, it was found that the rock had changed its shape by "flowage."

SUMMARY COMPARING DYNAMIC AND CONTACT METAMORPHISM

Contact Metamorphism	Dynamic Metamorphism
Intrusive rocks are situated near by.	No necessary connection with igneous rocks.
Areas affected are restricted.	Large areas are affected.
Many rocks showing contact metamorphism are not greatly folded.	Rocks showing dynamic metamorphism generally highly folded.
Recrystallization, generally with addition of considerable material.	Recrystallization, generally without addition of appreciable material.
Texture generally not banded except where rock altered was banded.	Texture generally schistose with parallel alignment of minerals.
Characteristic minerals: actinolite, wollastonite, tremolite, andalusite, garnet (andradite), etc. Many minerals common to both groups.	Characteristic minerals: muscovite, biotite, chlorite, wollastonite, actinolite, garnet, etc. Many minerals common to both groups.

Static, or Load, Metamorphism.—Metamorphic rocks may be formed without either deformation or igneous intrusion. If sedimentary rocks are deeply buried under a heavy load of overlying rocks, an increase of temperature results. With the rise of temperature and of pressure, the rocks are transformed without mass movement or deformation. The rocks resulting from such recrystallization differ from those produced by dynamic metamorphism. Dynamic action commonly destroys the original textures, whereas static metamorphism at certain places may allow such structures as cross-bedding to persist.

Static metamorphism may continue after the differential stresses due to deformation are relieved. Certain schists contain minerals that are not lined out parallel to the planes of schistosity. Large crystals called metacrysts (Fig. 429) are formed in schists lying across such planes. It is believed that these minerals were formed after the other minerals of the schist had developed and after the intense movements due to pressure had subsided. Pressure existed, to be sure, but it was the pressure of the overlying load of rock rather than the pressure and movement that are the conditions necessary for dynamic metamorphism.

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[1] *Jour. Geol.*, vol. 25, pp. 597–637, 1917.

STRUCTURES AND TEXTURES DUE TO METAMORPHISM

Rock Cleavage.—Cleavage is the property of a rock which causes it to split or break in certain directions more easily than in others. It is a secondary structure induced by the metamorphic processes. The cleavages of metamorphic rocks may be grouped as fracture cleavage and flow cleavage.

Fracture Cleavage.—Fracture cleavage is the property of a rock which causes it to break along closely spaced parallel fractures. It is independent of the arrangement of mineral grains. The fractures are due to shearing forces, and they may be open or closed or partly healed with soft cleavable minerals. If a rock has a single set of cleavage fractures, it will break into a series of thin sheets (Fig. 430); but if the fractures occur in two intersecting sets, the rock will break into polygonal forms.

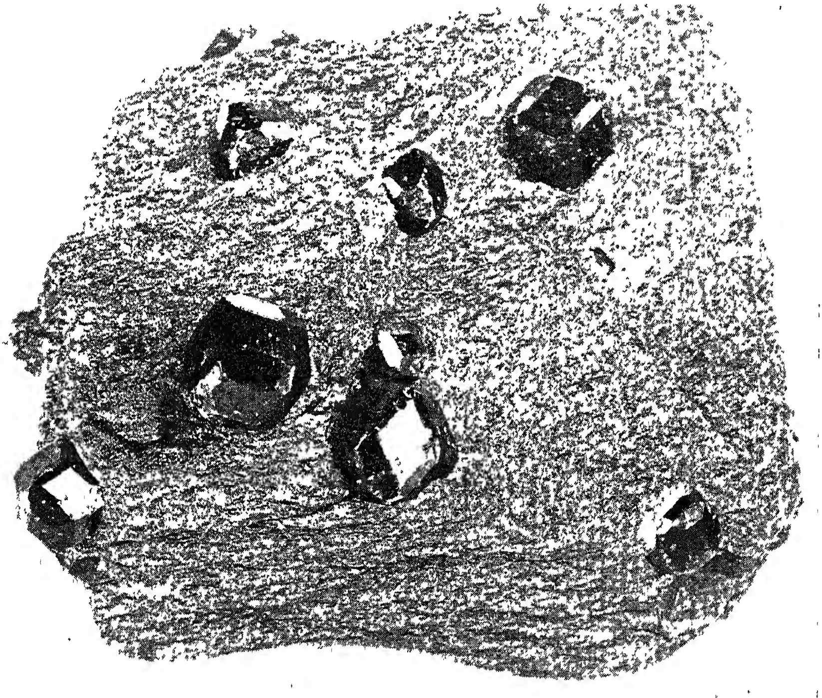

FIG. 429.—Garnet in schist, Stikeen River, Alaska. (*Courtesy American Museum of Natural History.*)

Flow Cleavage.—Flow cleavage is the result of "flowage" induced by dynamic forces. It is due largely to the presence of platy or needle-like minerals such as mica, chlorite, and amphibole, and the breaking takes place between the platy or needle-like minerals and along cleavages of the minerals themselves. If the minerals are arranged with their long dimensions parallel, the rock breaks so that the minerals separate on planes parallel to the long dimension of the grains. This type of cleavage is also called schistosity. It depends upon the parallel arrangement of mineral constituents.

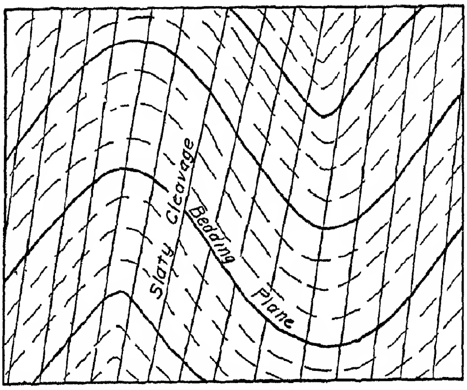

Fig. 430.—Folded beds of slate with slaty cleavage developed across the beds.

Development of Flow Cleavage.—If a mud or shale is examined under the microscope, it is found to consist of small particles of quartz which are mingled with finer particles of clay. Under great pressure the quartz grains are broken, and they are rotated so that their long axes lie in the direction of least pressure; the finer clayey material is recrystallized, and new minerals are formed such as mica and amphiboles.

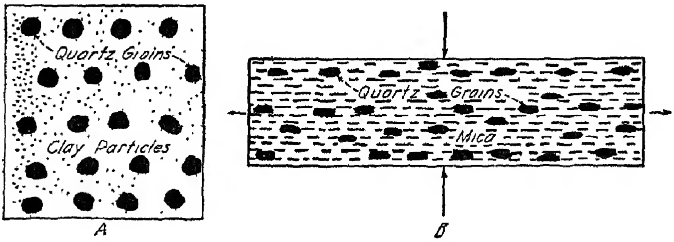

Fig. 431.—*A*, mud or shale greatly enlarged, consisting of grains of fine quartz sand and of smaller clay particles; *B*, the same mud or shale metamorphosed by pressure to form slate. The quartz grains are broken and the quartz fragments are oriented so that their long axes lie parallel to the direction of least pressure. Mica crystals form with flakes parallel to the direction of least pressure.

These new minerals have forms that are aligned with their long dimensions parallel to the direction of least pressure, as is illustrated by Fig. 431, in which *A* represents a shale greatly magnified and *B* shows the same rock after dynamic metamorphism. The direction of greatest pressure is indicated by the long arrows, and that of less pressure is shown by the short horizontal arrows. Because the pressure is less, the material of the rock tends to move out or to lengthen in that direction if free to move and to become shorter in the direction of greatest pressure. Because the mineral fragments are strung out in the direction at right angles to the lines of greatest pressure and because the flaky or fibrous minerals that are formed are aligned in that direction, the rock will break, or cleave, most readily in planes parallel to the long minerals. Thus a slaty cleavage is developed. Where bedded rocks are greatly compressed by folding, they take on a slaty cleavage, so that they often break across the beds, particularly at the axes of the folds. Not all slates are formed from muds and clays; basalts and other igneous rocks also may be converted into slates. Shales are very readily changed to schists, because they are weak and readily "flow" and also because they usually contain aluminum, potash, and iron, from which flaky minerals may form. If much potash is present, mica usually will be abundant in the product, and mica is among the most prominent platy minerals in the schists. Mica schists are common metamorphic products. Amphiboles are formed from calcium, iron, aluminum, and silica, and the amphibole schists and slates also are common products. Pure limestone and pure quartzite, on the other hand, do not so readily form schists, because they do not contain the materials for the formation of the platy minerals.

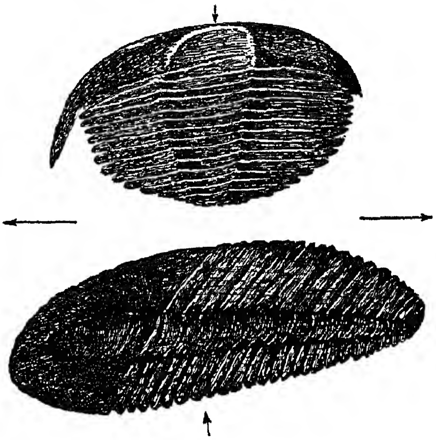

Fig. 432.—Drawing of fossils in a rock containing two trilobites that have been deformed by pressure. Originally the fossils had about the same shape. Short arrows indicate direction of shortening, and long arrows show direction of elongation of the rock and the contained fossils. (*Redrawn from a drawing of a specimen in the Marburg Museum as presented by Kayser, in Lehrbuch der Geologie.*)

When slates or schists are formed from shales or other rocks, there is a noteworthy lengthening in one direction and a shortening in the other. This is implied by Fig. 432, where elongation of fossils parallel to the cleavage in shale is shown.

Relation of Cleavage to Bedding.—If a series of rocks is thrown into close folds, the rock layers bend at the crests of the folds, and as a rule small folds are formed between the larger ones. During folding, the material of the beds changes to meet the new conditions, after the manner already described. Slaty cleavage is developed, and along the limbs of strong folds the cleavage tends to follow the beds. The cleavage, however, is essentially parallel over large areas, and at the crests of folds it cuts the beds nearly at right angles.

Age of Crystalline Schists.—The conditions favoring dynamic metamorphism suggest that the process is most active at great depth in the "zone of flowage." Except in very weak rocks, these conditions rarely, if at all, have prevailed near the surface during any geologic epoch. They may have been brought about, possibly within a few thousand feet of the surface, by magmas rising into the lithosphere, but where such conditions have prevailed the changes have been local and periodic. It follows therefore that the areas that contain the schists are areas where deep burial has taken place and where mountain-making movements have occurred. These were followed by long periods of erosion. Nearly all of the most ancient rocks of the earth's crust are appreciably recrystallized. This suggests that they have been most deeply buried. It is possible, however, that the outer part of the earth was generally hotter during early history than later. If a molten zone once existed below the surface, rock flowage and recrystallization could have taken place nearer the surface. Not all crystalline schists are geologically ancient. At certain places relatively young rocks have been converted into schists, and subsequently these have been exposed by erosion.

Metamorphic Rocks.—All metamorphic rocks were originally igneous or sedimentary rocks, but in some of them the original rock has been so greatly changed that its original nature may not easily be recognized. Consequently it is practicable to name a third class, the metamorphic rocks, to include them. Certain minerals that are found in igneous or in sedimentary rocks are very common also in metamorphic rocks. These include quartz, the feldspars, muscovite, biotite, amphiboles, magnetite, etc. Other minerals are found chiefly in metamorphic rocks. These include garnet, staurolite, talc, chlorite, serpentine, and many others. Many of these minerals are relatively heavy. When the rocks are under pressure, the formation of these heavy minerals tends to reduce volume. The conditions are the opposite of those where rocks alter at the surface of the earth by weathering, where the volume of the rock material is generally increased by the processes of weathering and decay.

Most metamorphic rocks are more or less foliated (Latin, *folium*, a leaf) or arranged in bands. This texture is due to the parallel arrangement of their constituent minerals or to the elongation of bands, layers, or lenses of granular material. The rocks will split more readily on planes parallel to the planes of foliation than across them. Often these planes are highly irregular and undulating. In a coarse-grained metamorphic rock like a gneiss the planes are far apart. In a fine-grained rock like slate or schist they are closer together.

Metamorphic rocks are classified, on the basis of their structure and texture, into foliated and non-foliated groups. The foliated rocks show a slaty, schistose, or gneissic structure, whereas the non-foliated rocks commonly are massive.

Slates.—A slate is a homogeneous, fine-grained rock which will split into thin or thick sheets with relatively smooth surfaces. The chief minerals are not readily distinguished by the naked eye. The surfaces are smoother than those of schists because of the finer grain. Some slates, however, have crumpled and folded cleavage surfaces. In some slates cleavage is parallel to the bedding, but at many places it intersects the bedding at high angles (Fig. 430). The original bedding planes may appear as streaks, often more or less plicated and running at any angle with the slaty cleavage.

Slates range in color from gray through red, green, and purple to black. The gray and black colors generally are due to carbonaceous material in the original rock, the carbon compounds having changed to graphite. The red and purple shades are due to iron and manganese oxides, and the green to ferrous iron silicates. Slates commonly are named from the predominant color or most conspicuous mineral. There is no sharp boundary between slates and shales or between slates and phyllites.

Phyllites.—A phyllite is a foliated, finely micaceous rock, of nearly uniform composition. It is coarser and more lustrous than slate but too fine grained to be classed as a schist. In phyllites generally the mica flakes are large enough to be distinguished with the unaided eye, but most of the other material is very fine grained. Phyllites represent a degree of metamorphism greater than that for slate but less than for schist. They may grade into either of these rocks.

Schists.—Schists include foliated metamorphic rocks which split readily into thin flaky slabs or sheets. The mineral grains generally are sufficiently large to be identified with the naked eye. The individual leaves are not of uniform thickness but are flattened lenses that often are bent and curving, with their platy surfaces in parallel planes. At right angles to these planes, schists break with much greater difficulty, leaving irregular frayed edges.

SOME COMMON MINERALS OF CRYSTALLINE SCHISTS

Generally without marked elongation	Generally elongated in two dimensions, platy or tabular	Generally elongated in one dimension, needle-like
Garnet	Muscovite	Actinolite
Pyrite	Biotite	Tremolite
Magnetite	Chlorite	Hornblende

Schists generally are classified on the basis of their mineral composition. If micas are prominently developed, the rock is *mica schist.* If hornblende is the mineral responsible for the foliation, the rock, which is generally dark green with a silky luster, is *hornblende schist.* Schists rarely have abundant feldspar. Commonly the various schists are formed from shales, but some are formed by the metamorphism of fine-grained igneous rocks such as felsites and basalts.

Gneisses.—A gneiss is a banded, coarsely textured metamorphic rock with a rough foliation developed generally as a result of dynamic metamorphism (Fig. 433). The alternating bands or layers are commonly of unlike mineral composition. In most gneisses feldspar is a prominent constituent; often the presence of large feldspar crystals serves to distinguish a gneiss from a schist. Gneisses contain the same minerals that are found in granitoid igneous rocks.

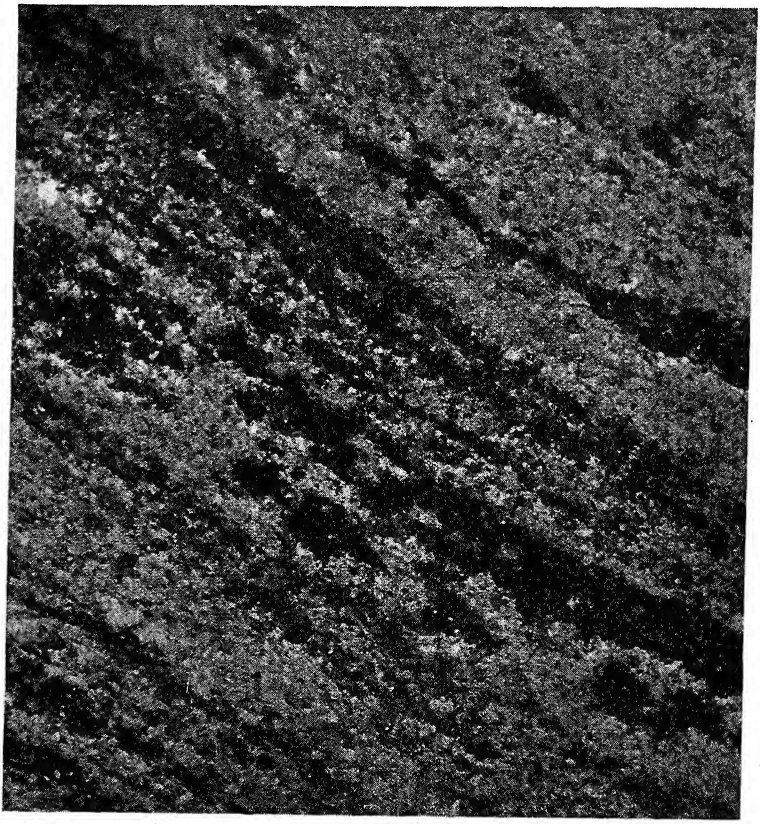

FIG. 433.—Granite gneiss (natural size). (*Courtesy Security Printing Company, St. Cloud, Minnesota.*)

Some gneisses are banded when first solidified. Such texture is called *flow banding,* and the rock is a *primary* gneiss. Certain crystals formed before the magma came to rest, and the layering is due to movement of the partly liquid mass. Recent work has shown that primary gneisses are more common than was formerly supposed. Where banded gneiss intrudes vesicular basalts in which the vesicles are still spherical and not deformed, it is a rational inference that the gneiss is primary.

Granites are changed to gneisses by dynamic metamorphism. During movement under great pressure the crystals are rotated so that they lie with their long axes lined out in the direction of least pressure (Fig. 433). Recrystallization of micas takes place also, and these lie with their leaves parallel to the long axes of the rotated crystals. The changes are brought about less readily than the change from shale to slate, because the granite is a stronger rock, yet at great depths the granite acts as a somewhat plastic mass, and its minerals rotate much as the minerals rotate when schist is formed. Gneisses have formed also from arkose sands by granulation and cementation of fragments of feldspar and other minerals which make up the sands. Certain granite gneisses appear to have formed by movement during the consolidation of the granite probably just after the granite had cooled to a pasty mass but had not become completely solid. In such granites the feldspar and mica crystals are drawn out and arranged in lines so that the structure of the gneiss is much like that of the granite gneiss formed by dynamic metamorphism.

In the following table the chief sedimentary rocks are named: (1) as they are laid down, (2) after consolidation, (3) after metamorphism. Another table shows the igneous rocks and their metamorphic equivalents.

SEDIMENTARY ROCKS AND THEIR METAMORPHOSED EQUIVALENTS

Unconsolidated	Consolidated	Metamorphosed
Gravel	Conglomerate	Conglomerate schist
Sand	Sandstone	Quartzite, quartz schist
Mud, clay	Shale	Slate, mica schist
Calcareous ooze	Limestone	Marble, calcareous schist
Peat	Lignite, bituminous coal	Anthracite coal, graphite

IGNEOUS ROCKS AND THEIR METAMORPHOSED EQUIVALENTS

Igneous Rocks	Metamorphosed Equivalents
Granite	Granite gneiss
Syenite	Syenite gneiss
Diorite	Diorite gneiss
Gabbro	Gabbro gneiss
Rhyolite	Mica schist
Andesite	Hornblende schist
Basalt	Slate, hornblende schist, biotite schist, chlorite schist

Other Metamorphic Rocks.—The metamorphic rocks that do not show foliation such as is shown by gneisses and schists are *non-foliated.* A quartzite is a metamorphosed sandstone cemented by silica. If it is dynamically metamorphosed, it becomes a quartz schist. A *marble* which is coarse-grained recrystallized limestone, if metamorphosed by pressure so that it has a prominent cleavage, is a calcareous schist. *Soapstone* is a rock composed essentially of talc but commonly containing some mica, tremolite, or quartz. Most soapstones are metamorphic rocks that have been partly hydrated.

Metamorphism of Coal.—Coal is formed by the consolidation and induration of plant remains and largely by the induration of peat. There is a series: (1) peat, (2) lignite, (3) bituminous coal, (4) anthracite coal, (5) graphite. Each member of this series contains more carbon and less gas and water than the member preceding it. At many places coal-bearing formations may be followed from the plains, where they lie nearly flat, to mountains, where they are highly folded. As the amount of folding increases, the character of the coal changes. In the western plains, where the beds lie nearly horizontal, lignite only is present; toward the mountains the lignite has changed to bituminous coal. Where the coal beds have been more intensely folded, the coal by metamorphism has become anthracite.

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CHAPTER XVI

PROBABLE CONDITIONS WITHIN THE EARTH

Terrestrial Magnetism.—There is a certain variety of the mineral magnetite that possesses the property of drawing to itself bits of iron and steel. This mineral is known as *lodestone* (Fig. 434); and if a bar of steel is stroked with a lodestone, the bar also will possess the magnetic property. If it is placed in iron filings, these filings will be attracted to the bar and will be attached chiefly to the ends or poles of the bar (Fig. 435). Such a bar of magnetized iron is a *bar magnet.* The earth may be regarded as a gigantic bar magnet. If a magnetized needle is mounted so that it is balanced on a pin, it will be acted upon by the earth's magnetism, orient itself parallel to the earth's magnetic field, and point to the earth's magnetic pole. A needle so mounted is known as a magnetic compass, and it is used to point directions on land and on sea. At the magnetic poles the magnetic needle points directly downward. The magnetic poles are not points but small areas. The earth's magnetic poles do not coincide with the north and south axial poles. The magnetic pole in the northern hemisphere is located among the islands of northern Canada (Fig. 436) and is approximately at 70° 5′ N. and 96° 46′ W., and the opposite magnetic pole at about 72° 40′ S. and 152° 30′ E.

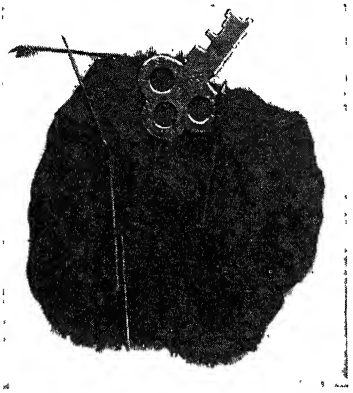

Fig. 434.—Photograph of a lodestone—a fragment of magnetite that attracts iron.

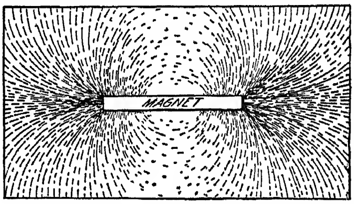

Fig. 435.—Sketch showing the arrangement of iron filings near a magnet.

In the magnetic compass the needle is mounted to swing horizontally, and the compass box is graduated in a 360-degree circle. Other needles, called dip needles, are mounted to swing in a vertical arc, and this arc also is graduated to read 360 degrees. Readings on both compass and dip needles are different at different places over the earth's surface, but some of these differences vary according to certain laws, and the variations of the compass in general are well known (Fig. 437), so that when

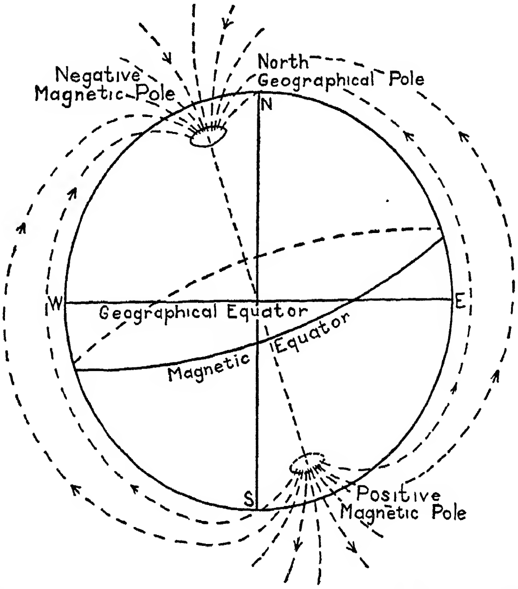

FIG. 436.—Diagram showing the relative positions of the magnetic and planetary poles of the earth. (*Based on a figure by Black and Davis.*)

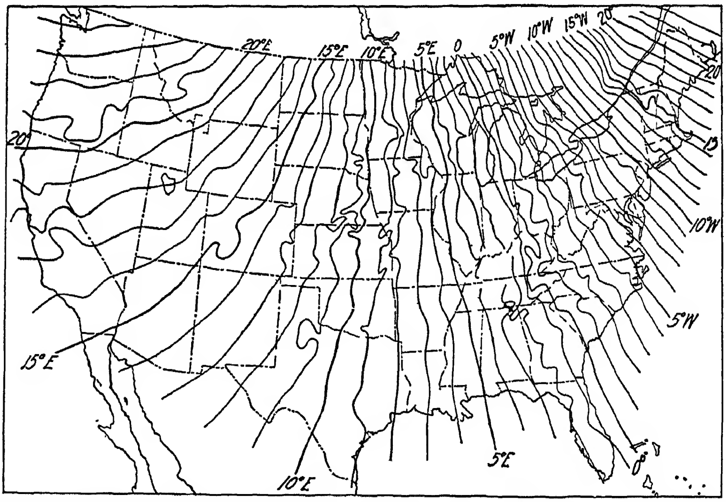

FIG. 437.—Isogonal lines showing the declination of the magnetic needle throughout the United States. East of the zero line the north end of the compass needle points to the west of north; west of that line it points east of north. (*After U. S. Coast and Geodetic Survey.*)

it is used for finding directions, allowances are made for variations which result in part from differences in the earth's geographic and magnetic poles.[1] The magnetic needle is a very useful instrument for the mariner and for the geologist, although at places it is unreliable owing to local magnetic attraction. The dip needle shows great variations. Thus a needle will read different intervals on its arc on different days and at different times during the same day. The dip needle is used for locating magnetic masses near the surface of the earth. Some of these magnetic bodies are valuable ores. Certain beds that contain small amounts of magnetic substance may be followed by dip-needle readings, and by using these the positions of the various rocks may be discovered and the geologic structure worked out in an area even where the beds at most places are concealed by mantle rock.

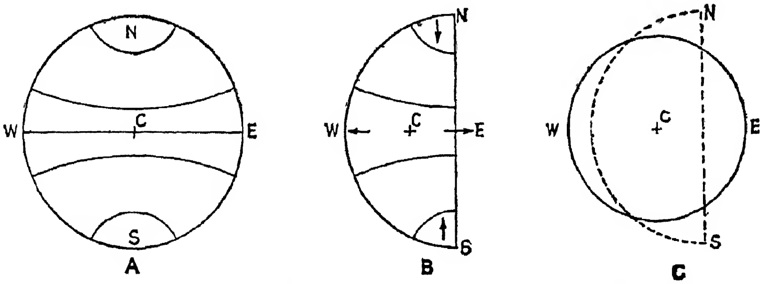

Fig. 438.—Diagrams illustrating isostasy. If the earth *A* were split into hemispheres the hemisphere *B*, would not retain the shape of *B* but by isostatic adjustment would assume the shape of sphere *C*. Each of the small spheres would have great circles equivalent to that shown by the solid line in *C*.

Theory of Isostasy.—Most great heavenly bodies are essentially spherical units. If one were of another shape, it would soon assume the nearly spherical shape by readjustment of its material. If the earth for some reason were cleaved through its center and the two parts separated so that each became an independent body, the two halves quickly would become essentially spherical. This is illustrated by Fig. 438. The center of gravity of the hemisphere is *C*, and the polar regions are *N* and *S*. The material at *N* and *S* is drawn by gravity toward *C*. The rock matter is so weak that it cannot hold against the great pressures due to the weight of the rocks. Material at *N* and *S* would move toward *C*, and material at *E* and *W* would be pressed out by the force of the material of *N* and *S* moving toward the center of the mass. The body would tend to become spherical if it were homogeneous; but if part of it near the surface were heavier than other parts, it would be drawn down farther by the force of gravity, and the place above it would be

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[1] Poles of bar magnets attract unlike poles and repel similar poles. The positive, or +, end of a bar magnet is sometimes called the north end because it points north and is attracted by the magnetic pole in the northern hemisphere. That pole is the negative, or −, magnetic pole of the earth, corresponding to the −, or "south," end of a bar magnet.

nearer the center of the sphere than any other place at the surface. Similarly if another part of the sphere were lighter than the remainder of the sphere, it would not be drawn so near the center of the mass, and the surface of the sphere above it would be higher than other parts of the sphere at the surface. Small bodies of strong material would not tend to become round, because their pressures would not be sufficient to crush their material.

That the earth is a yielding body is shown by the earth tides. The rock sphere responds to the attraction of the moon and the sun just as the ocean does (page 234). Although the rock tides are much smaller than ocean tides, they have been measured and studied.

According to the theory of isostasy, the earth's great sectors are of nearly equal weight. The great mountain systems and the plateaus are believed to be made up of and underlain by light rock, and the ocean basins are believed to be underlain by heavy rock, so that they are almost in balance, as if the masses were floated on a liquid. This theory is based on physical observations: (1) on plumb-bob deflections and (2) on gravity determinations.

If a plumb bob is hung from a string suspended above a great, flat plain, it points approximately toward the center of the earth. This is due to gravity or the attraction of the earth. If the plumb bob is hung above a plain near a great mountain range, it is attracted also by the mountain range and is deflected or drawn toward it. Calculations show, however, that the plumb bob is not deflected toward the mountains so much as one would be warranted to expect from their size. There is evidently a deficiency of mass due to lighter material in or below the mountains.

The pendulum also may be used for measuring the pull of gravity at various places on the earth's surface. The period of a pendulum of a given length depends upon the pull of gravity, which varies with the mass of material below the pendulum. The stronger the pull the faster the pendulum vibrates. The vibrations are recorded by a clock. If the pull is abnormally strong, the pendulum vibrates faster and the clock gains time; and if the pull is abnormally weak, the pendulum vibrates more slowly and the clock loses time. By allowing the clock to run over long periods, even small differences of the pull of gravity can be measured. Stations have been established at many places over the earth's surface, and observations made. These observations show that the mountain sectors are heavier than the sea-level sectors, but they are not so much heavier as they are larger.

The theory of isostatic compensation is illustrated by Fig. 439. The three columns represent three prisms of the earth. In one the surface of the ground is above sea, in another it is at sea level, and in the third it is below sea. The three prisms are of equal mass and weight; the longest is made of the lightest material, and the shortest is of the heaviest material. The line on which the blocks rest is called the zone of compensation. It is a theoretical plane or zone, and the rock columns above it are supposed to be of equal weight. Below the bases of the three columns the earth is solid, but, because of the great pressures, it is believed to yield easily as a viscous liquid does. This conception of isostasy is accepted more generally by other students of the earth than by students of geology.

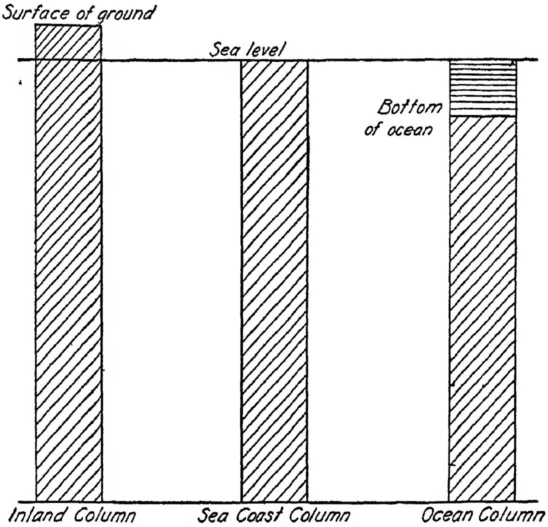

Fig. 439.—Diagrams showing three columns of rock, one extending above sea level, another to sea level, and a third to the bottom of the sea. According to the interpretation of isostatic compensation the weight of the first column equals that of the second and also that of the third column plus the weight of the column of water above the third column. (*After Hayford and Bowie.*)

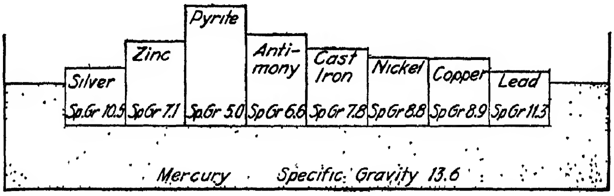

Fig. 440.—Ideal illustration of the isostatic assumption that differences in altitude of large parts of the earth's crust are compensated by differences in density. The blocks are equal in cross section and in weight and thus sink to equal depth. (*After Bowie.*)

An illustration of an isostatic condition is shown by Fig. 440. Blocks of various materials of different densities are immersed in quicksilver, which has a density 13.6 times that of water (sp. gr. 13.6+). The blocks weigh the same and have the same cross sections but are of different lengths. The densities of the materials are shown on the blocks. The blocks of the lighter materials extend higher above the surface of the mercury than those of heavier materials. In Fig. 441 all of the blocks are of copper (sp. gr. 8.9) but are of different lengths. The longer ones sink deeper and also extend farther upward than the short ones. If material is taken from the top of a high block and is added to a low one, the base of the high block will rise and the base of the low one will sink. Adjustment is attended by the movement of the mercury from the bottom of the depressed block to the bottom of the elevated block. It has been stated that the erosion of a mountain range may cause the mountain range to rise as pressure due to the weight of the eroded material is relieved, and the block below the basin upon which the eroded material is deposited will sink. This process has been invoked to explain deep basins in which very great thicknesses of shallow-water sediments have been deposited. As the basin was filled, it sank, and it continued to afford basin conditions for deposition. As the land was eroded, it was pushed up, and it continued to afford conditions for erosion. It is believed that certain basins have sunk while they were being filled (see Deltas) and that land has risen slowly to supply a source of new sediments (page 380).

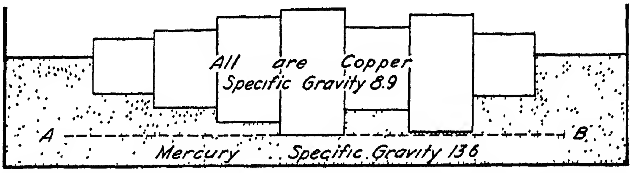

Fig. 441.—Diagram showing seven copper blocks of equal cross sections but of different heights; all are immersed in mercury. The longest blocks rise highest and also extend to the greatest depths. According to the theory of isostasy, higher mountains are underlain by lighter material than low ones and both are in balance. (*After C. R. Longwell.*)

The mechanism of this phase of isostatic adjustment may be illustrated by the U tube. Suppose that the tube (Fig. 442) contains water which rises in the tube to the level indicated. If water is poured in one arm, it rises in both arms to the same level. Now suppose that a wooden cube half as heavy as water with a 2-inch edge and with a base having half the area of the cross section of the tube is dropped in the arm *A*. The water will rise to the same level in both arms, and one-half of the cubic block will be submerged. Suppose that a cube is dropped in each arm. Each cube will be half submerged. If one cube is removed from arm *B* and placed in arm *A*, the water rises to the top of the lower cube in arm *A* (Fig. 443) and flows from *A* to *B*. The movement of water is supposed to illustrate the movement of rock in accordance with the demands of the theory of isostasy.

An alternate theory is that the earth near the surface acts like a steel spring (Fig. 444). It is strong but elastic. If material is taken from *A* and added to *B*, the section *B* will be depressed or sunk, and owing to relief of pressure *A* will rise. In an area of deposition *B*, loading causes contraction. In this figure there is no deep-seated movement of material from *B* to *A*, as there is in the U tube. In some respects this theory seems best to accord with the conditions within the earth and to accord with the strength of the earth. It is certain that the earth can accumulate strains over long eons of time. It tends to assume a shape in accord with the theory of isostasy, but imperfectly so. The atmosphere is isostatic; so is the hydrosphere; but the lithosphere is possessed of great strength and is rigid, as is shown by the study of earthquake waves.

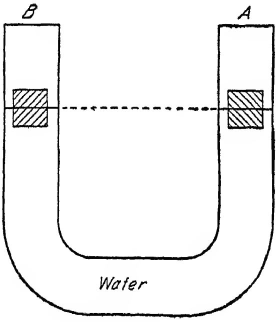

FIG. 442.

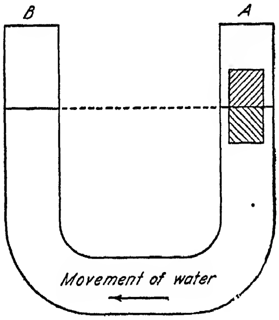

FIG. 443.

FIG. 442.—Diagram showing a U-tube partly filled with water. A cube of wood half as heavy as water and with sides half as large as the cross section of the tube is placed in each arm. The blocks rise to equal elevations.

FIG. 443.—The block is moved from arm *B* and placed in arm *A*. The water level will be at the same elevation as in Fig. 442. There will be a movement of water from *A* to *B*, to compensate for the movement of the block from *B* to *A*. This illustrates one interpretation of isostatic adjustment.

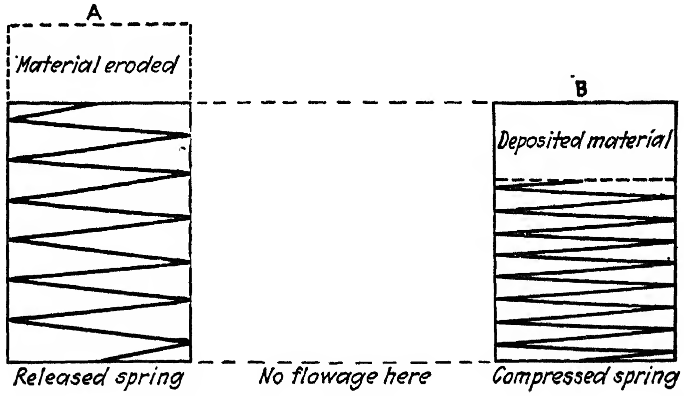

FIG. 444.—Diagram illustrating how a rigid yet elastic earth body may respond to removal of load (column *A*) and to the deposition of load (column *B*). The elastic earth acts like a steel spring. Column *A*, due to removal of load by erosion, is relieved of weight and tends to rise. Column *B* is compressed by the weight of material deposited and tends to sink. Flowage of rock from the base of *B* to the base of *A* is not necessary for adjustment, as in the case with the U-tube experiment. (*Based on a discussion by Bailey Willis.*)

It does not "flow" as water and mercury flow in the experiments cited, but it flows as an incompetent solid by adjustments due to shearing, to recrystallization, and to the formation of small fractures which are soon closed by pressure; also by the granulation of rocks and by other changes that take place during dynamic metamorphism. It accumulates stresses through long periods without important readjustments, as is shown by peneplanes, which are developed during erosion. Peneplanes are common features. They have formed again and again in the geologic past. If the earth's surface responded quickly and perfectly to changes of load, the peneplane could not readily form. As soon as a mass of rock was eroded from a high area, that area would rise again, and it would be subjected to renewed erosion. The earth responds to great load stresses by adjustment but probably not to the smaller ones.

The Earth's Interior.—The crust of the earth, as already stated, is the part of the solid earth that comes under observation of man. Rocks once buried as deep as 5 or 10 miles have been exposed by erosion, and they may be studied, but little is known of the nature of the earth below a depth of 5 to 10 miles. Parts of the crust of the earth may be weighed and analyzed, and experiments may be made with them. Below the crust one passes from the region of observation to the realm of speculation. There are nevertheless many data which have a bearing on the conditions that exist within the earth, and by use of these something may be learned of its character.

It is known that the interior of the earth is hot, for hot materials are expressed from it. Thermometer measurements of the temperatures in mines and bores show that the temperature of the earth's crust increases steadily downward. In regions of hot springs the increase is as high as 1°F. per 30 or 40 feet. In mines that have been opened for some time, where the rock is cooled by air circulation, the increase with depth is much less rapid. The average increase is about 1°F. per 60 or 65 feet, or approximately 1°C. per 100 feet.

If this increase continued to a depth of 100 kilometers, or 60 miles, the temperature would be about 3000°C., which is far above the melting points of ordinary rocks at the earth's surface, but because of pressure the rocks are mainly in a solid state. Very little is known of the temperatures within the earth. It is improbable that they increase downward at a uniform rate. It is certain that most of the earth is essentially solid, for most of it transmits earthquake waves of a character that do not pass through liquids. If, by compression or by radioactivity, parts of the earth become molten, it is probable that some of this hot material rises toward the surface and carries heat with it. It is not unlikely that the outer part of the earth has a relatively high temperature as a result of this process.

The pressures within the earth are very high owing to the weight of rocks. The densities of rocks are known, and these pressures may be calculated. At the center of the earth the pressure is more than 3,000,000 atmospheres.

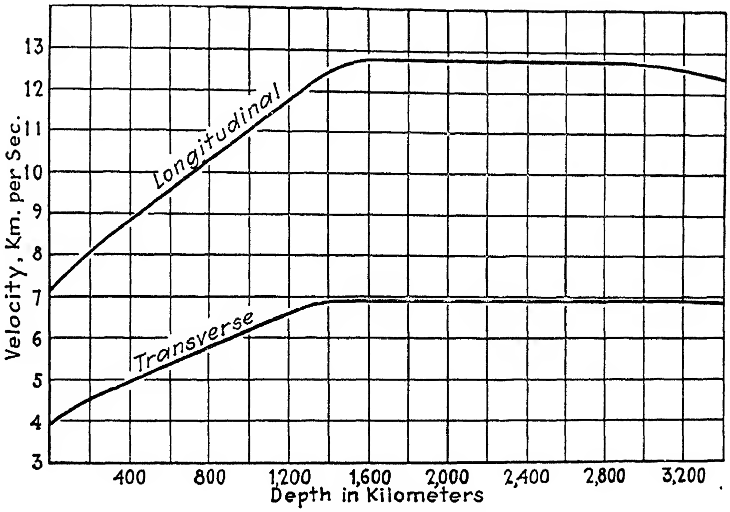

FIG. 445.—Curves showing the velocities of longitudinal and transverse earthquake waves at various depths below the surface of the earth, as calculated from seismograms. (One kilometer is 0.62 mile.) The speed of the longitudinal or compressive wave is everywhere about 80 per cent faster than that of the transverse wave. (*After Adams and Williamson.*)

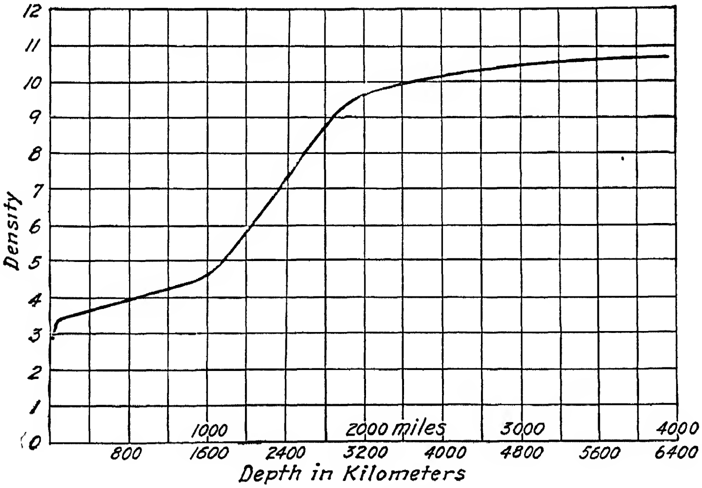

FIG. 446.—Diagram showing the density of the earth at various depths. (*According to estimates of Adams and Williamson.*)

Earthquake shocks are transmitted from their points of origin around the earth and also through the earth. The waves that pass around the earth move at a rate of about 2 miles per second. Those that travel through the earth are of two kinds: one, called transverse waves, travels between 2.5 and 4.2 miles per second; and the other, called longitudinal waves, travels between 4 and 8 miles per second. The rates at which waves travel depend upon the elasticity of the material through which they pass. The greater speed is made through the most elastic rocks. Seismographs are set up and are in continuous operation at many places over the earth, and they record the time of arrival of the earthquake waves at many places. Thus if there is an earthquake at *A* (Fig. 345), a station at *B* will record (1) the time of arrival of the fast longitudinal wave, (2) the time of arrival of the slower transverse wave, and (3) the time of arrival of the wave that passes around the surface of the earth. Other seismographs record arrivals at other places (Fig. 346).

The rate of travel of earthquake waves has been calculated for many earthquakes, and it is found that both the longitudinal and the transverse waves move along lines passing through an outer shell of the earth that has a thickness of 2,000 miles, or half the radius of the earth. Since the transverse wave does not pass through liquids, and since it does pass through this shell, it is inferred that the outer shell of the earth having a thickness of 2,000 miles is essentially solid. The volume of this shell is about seven-eighths that of the earth. The core that it surrounds has a radius of one-half that of the earth and a volume of about one-eighth that of the earth. Since the transverse wave is not known to pass through it, the state of the center is not known.

The velocities at which earthquake waves travel at various depths are shown by Fig. 445.[1] The velocity, as stated, depends upon the elastic properties and the density of the material through which the waves pass. Since the speeds of the waves shown by Fig. 445 are related to density, an estimate of the density of the earth at various depths may be made.[2] This has been done by Adams and Williamson, and the results of their calculations are shown by Fig. 446. The density of rocks at the surface is known to be 2.7 times, and that of the entire earth 5.52 times, that of water. The density at the center of the earth is estimated to be 10.7. If the earth were made of the same material from the surface downward, its density should vary with the pressure, and presumably its elasticity and rigidity would vary approximately with the pressure. This evidently is not the case, however, for the rate at which earthquake waves are transmitted changes very irregularly with depth (Fig. 445). At a depth of 60 miles the rate of increase of speed falls off, and at a depth of about 1,600 kilometers (or 1,000 miles) the velocity of earthquake waves becomes about uniform.

[1] ADAMS, L. H., and E. D. WILLIAMSON, The Composition of the Earth's Interior, Smithsonian Inst. *Ann. Rept.* 1923, pp. 241–260, 1925.

[2] Certain other assumptions enter into this calculation.

If the earth were composed of the same material from the surface downward, for example of granite, it should be expected to show a nearly uniform change in density from the surface downward; because it does not do so, it is reasonable to suppose that the nature of the material changes, and it is supposed that the earth is made up of zones of material of different composition. The nature of these zones, according to estimates of Adams and Williamson, is stated below (see Fig. 447).

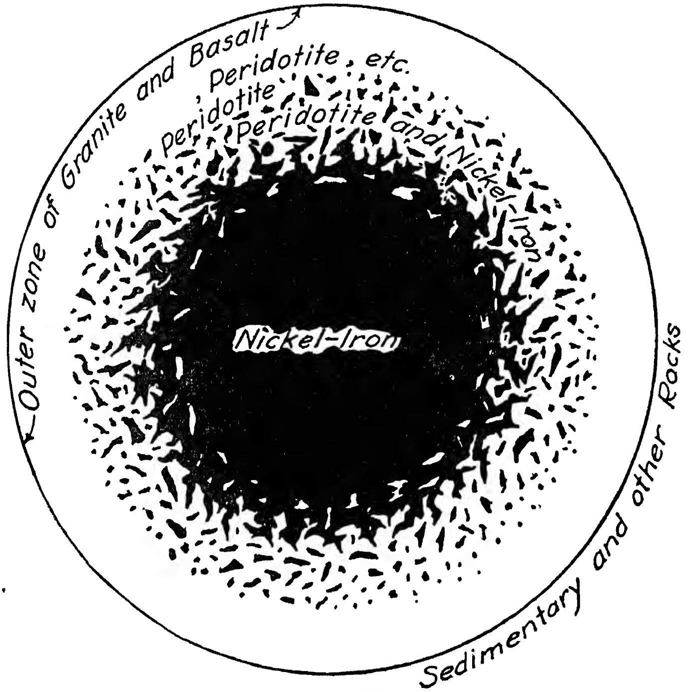

FIG. 447.—Schematic diagram showing the composition of the earth's interior, as estimated by Adams and Williamson.

1. The outer shell is one composed chiefly of sedimentary rocks. At places this shell is wanting, and granite or gneiss or other igneous rocks crop out at the surface. At other places the sedimentary shell is known to be about 2 miles thick. Its average thickness is probably between 1 and 2 miles.

2. At many places granite or granite gneiss is found below the sedimentary shell. The rigidity of the earth increases rapidly to a depth of 60 kilometers, or about 37 miles. This zone with a thickness of about 37 miles under continents is believed to be made up chiefly of granite and gneiss but becomes more basic, probably changing to gabbro near its base.

3. Below the granite-gneiss shell is one of basic rocks, probably peridotite in the main, which contains much iron in silicate minerals

but probably no iron in the metallic state. In this material the earthquake wave travels at a velocity increasing with depth until it reaches a zone about 1,600 kilometers, or about 1,000 miles below the surface (Fig. 445).

4. Beyond the depth of 1,000 miles the velocity of the earthquake wave is about uniform. Although it passes through more highly compressed material, there is no increase in its speed, and it is believed that there is a change in the character of the material, so that the speed of the wave remains about uniform, notwithstanding the increase in state of compression of the material through which it passes. In this zone the material is believed to be pallasite, a mixture of the minerals of basic rocks and metallic iron, the latter increasing with depth. Probably increased pressure does not cause the velocity of the wave to increase in iron so much as it does in basic rock. The effect of steady increase of pressure might be balanced by increase of iron if the proportion of iron steadily increased, so that the velocity of the waves would remain essentially uniform.

5. At a depth of about 2,000 miles the velocity of the earthquake wave decreases, although it is passing through more highly compressed material. It is believed that this change is due to another change in the character of the material through which the wave passes and that below that depth the material of the earth is essentially iron or nickel iron. Little is known of the movement of earthquake waves below 2,000 miles. Both waves appear to slow up, and the transverse one seems to be absorbed.

These hypotheses meet the requirements of the curve showing the velocities of the earthquake waves and of the density curve better than any other and are the most reasonable that have been proposed.

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CHAPTER XVII

MINERAL DEPOSITS

Introduction.—Mineral deposits are those portions of the lithosphere that are of value to man because they contain certain mineral substances, such as coal, oil, gas, metals, mineral fertilizers, and salts. These materials are used by essentially all civilized peoples and are necessary for modern industrial progress. Many of them were used by man before the beginning of the Christian era. The total value of the mineral production of the United States in 1937 was $5,440,000,000. The value of metals was about 26.5 per cent of that sum; of non-metals about 15.4 per cent; and of coal, oil, and gas about 57.4 per cent of the total.

Prospecting.—The distribution of mineral deposits is governed by definite geological laws. The deposits were formed at certain times and at certain places because the conditions were favorable for their formation. A great many mineral deposits crop out or are exposed at the surface. Coal deposits frequently were discovered by men who observed the broken bits of coal, mixed with earth, that mark the outcrops of coal beds. The discovery of certain oil fields has been due to the presence of asphalt or mineral pitch on the surface of the earth. The oil, seeping out of fractures or of porous beds, hardened to form asphalt; and where suitable places were drilled near by, the oil was found.

Nearly all of the deposits of metalliferous ores that were worked in ancient times cropped out. The ores of gold, silver, and iron are weathered at the surface, but in most of them the metals have remained at the very outcrops. Copper, on the other hand, usually is dissolved from the sulphide ores and carried away. Many of the valuable copper deposits carry iron or some other metal, but very little copper, at the surface. The laws of alteration of ores are understood just as are the laws of occurrence, and in weathering at the surface each metal behaves in its characteristic manner.

Geophysical Prospecting.—In ancient times many believed that deposits of water, metals, and other valuable substances could be found by the use of a so-called "divining rod." Even today there are people who believe that this instrument has value. A common form of the divining rod is simply a forked stick. A fork is held in each hand, and it is claimed that as the operator walks over a valuable deposit the third prong of the fork will point toward the deposit. There is no basis for belief in the divining rod, and tests have shown that it has no value.

There are certain methods of prospecting that have a sound scientific basis. The magnetic needle, properly balanced, will point to a body of magnetic iron ore that is not too deeply buried below the earth's surface.

Along the Gulf Coast in Texas and Louisiana occur great masses of salt, called "salt domes," that are concealed below the surface. Valuable deposits of oil are found in sands above and around the margins of many of these domes. Certain of these salt domes have been located by the "torsion balance." The weight of a given mass of salt is different from that of an equal mass of the rocks that surround the salt, and the torsion balance, by measuring accurately the gravitative pull of the mass, has aided in its location.

The seismograph is used extensively for the same purpose. A charge of dynamite is exploded in the area that is being prospected, and the earthquake waves are recorded on seismographs that are suitably situated. Because the earthquake waves travel faster in salt than in the beds of clays and sands that surround the salt dome, it is possible to calculate the paths of certain waves and to estimate the position of the salt dome from such calculations. Many rocks other than salt also are located by the seismograph.

In recent years electrical methods for prospecting for metals have been developed. If electrical impulses are sent through the earth, their propagation in areas where metalliferous deposits are present may be different from their behavior in regions where there are no metalliferous deposits. Certain metalliferous deposits have been discovered by these methods, but at places faults, fissures, or bodies of underground water have been mistaken for mineral lodes. Thus much work remains to be done before these methods of prospecting are perfected.

Coal.—Coal is carbonaceous mineral matter that generally is consolidated as a result of being buried in the earth. It is used as a fuel and also for making gas, coke, chemicals, and many other products. Coal is the altered product of vegetable tissue. When dead vegetable tissue is exposed to the atmosphere, it decomposes and is scattered. If it finds lodgment in water, the oxygen of the air is excluded partly, and the vegetable tissue is preserved and changed to peat. When peat is buried, it undergoes further changes; and if it is buried deeply, it is partly consolidated, water is pressed out, and certain chemical changes take place resulting in a loss of certain gases so that the peat is changed to low-grade coal, or lignite. If the beds containing the lignite are deeply buried or are folded (Fig. 448) under load, further changes take place, and the lignite is converted to bituminous coal; if stronger folding takes place under heavy load, the bituminous coal is converted to anthracite coal. Thus, as noted (see page 400), there is a series: (1) woody tissue, (2) peat, (3) lignite, (4) bituminous coal, (5) anthracite coal. In this series, in

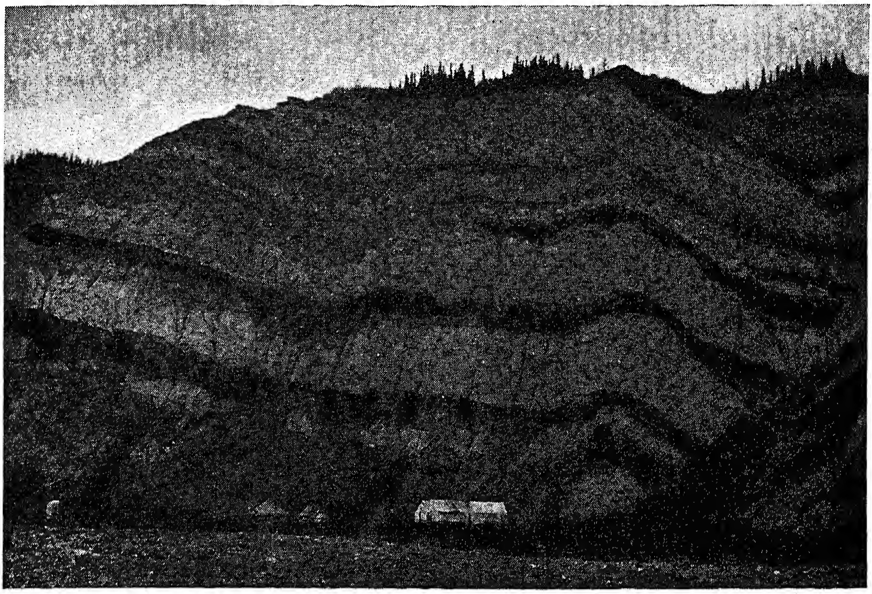

FIG. 448.—Strata of coal (black) interbedded with other sedimentary rocks along the valley of the Healy River, Alaska. (*Photograph by Evans.*)

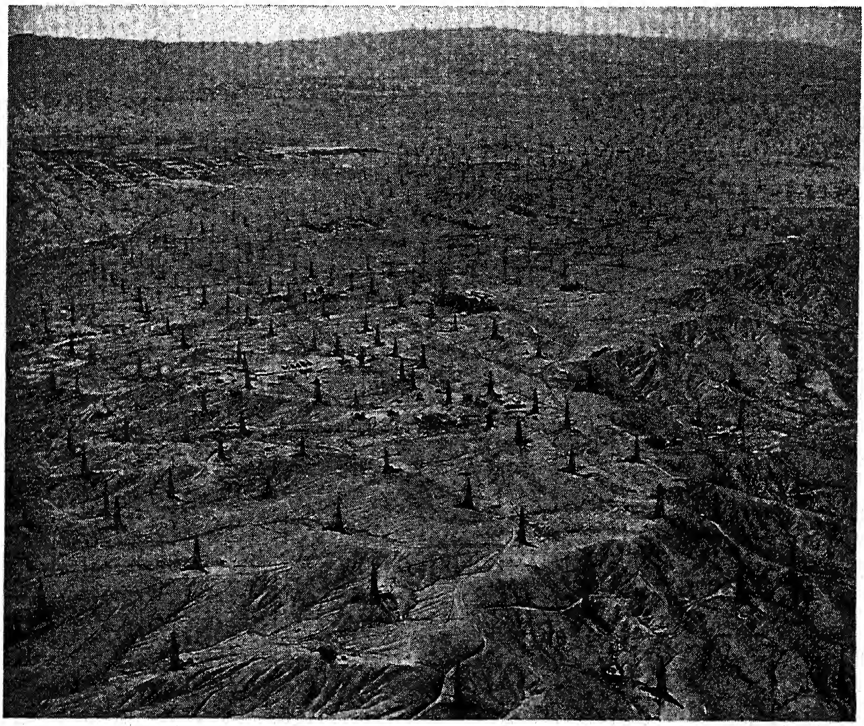

FIG. 449.—Oil field in Buena Vista Hills, California. (*Courtesy Spence Air Photos.*)

Fig. 450*A*.—Aerial photograph of the Medicine Bow structural dome, Wyoming. Arrows show direction of dip of beds. After folding the dome was deeply eroded. The broad black area is Medicine Bow river. Each small square is a square mile. (*Courtesy California Company.*)

the order named, water and gas decrease, which results in a proportionate increase of carbon and ash, although the total amount of carbon has been reduced. Anthracite coal is very high in carbon (85 per cent or more), and most of it contains considerable ash. It contains very little water or gas.

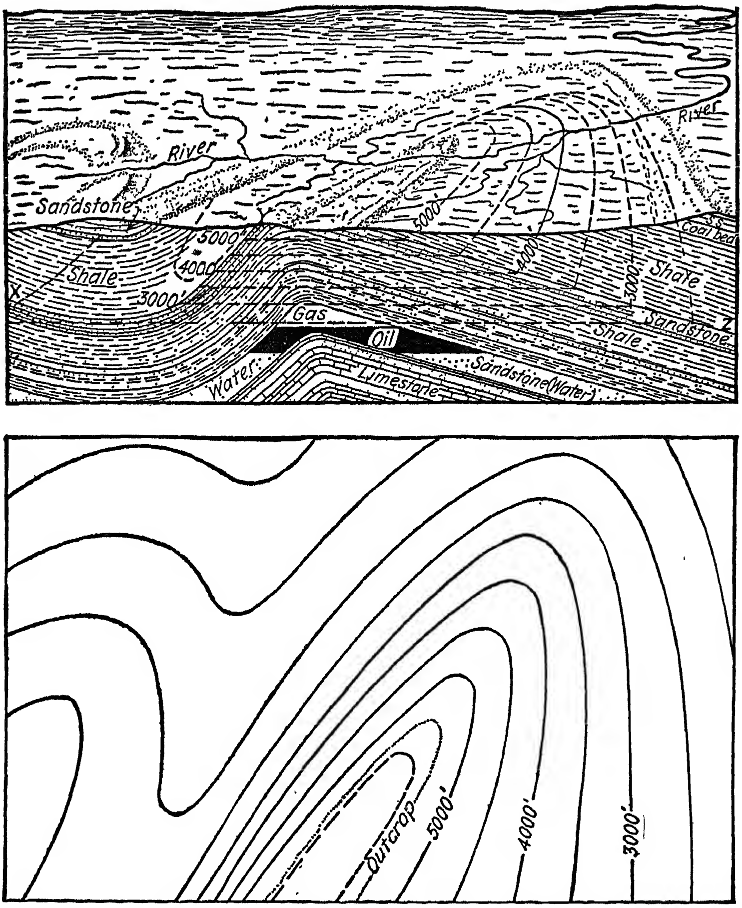

FIG. 450*B*.—Upper figure represents a cross section and sketch of an anticline, illustrating the use of structure contours and the occurrence of oil and gas. Lower figure is a structure contour map of the same anticline. The contours are drawn on top of sandstone *XYZ* and show approximately the position of this bed in all parts of area mapped. (*After Hewett and Lupton.*)

Petroleum.—Petroleum (rock oil) is an inflammable mixture of oily hydrocarbons that rises from the earth or is raised by pumping (Fig. 449). It is formed by the decomposition of plant and animal remains that have been buried in sedimentary beds. It is practically never found in igneous or in metamorphic rocks or in fresh-water sediments that are not associated with marine strata. Petroleum generally originates in muds, clays, shales, marls, or limestones, but it cannot ordinarily accumulate in shales in large amounts because suitable openings are not available. As a rule, it accumulates in sandstone or in porous limestone or dolomite.

Natural gas and salt water are nearly always associated with petroleum. It is the pressure of the gas that makes oil rise in wells and issue from gushers, or flowing wells. The salt water is believed to be sea water that filled the openings in the rocks when the sediments were laid down. Most oil-field waters, however, do not resemble sea water very closely in composition, and it is believed that the ancient sea water has changed by reacting with rocks and with petroleum during the long ages when it was stored with oil in the rocks.

Petroleum, along with gas, moves from shales to the porous rocks where larger openings are available. The muds and shales not only supply source beds for the oil, but generally they form a cover above the oil-bearing bed that prevents the escape of the oil. The porous rocks contain the salt water, gas, and oil, and the three fluids by gravity are separated so that the oil rises above the water and the gas above the oil. Nearly all oil-bearing strata have been thrown into folds by diastrophism, and many of the greatest oil fields of the world are found on anticlines like those shown by Figs. 450*A* and 450*B*. The gas occupies the central part of the fold; the oil lies below the gas on each flank of the fold; and salt water is found lower down on each flank. Not all of the oil fields are on anticlines. Many valuable ones are at unconformities or in sands that play out up dip. The great East Texas oil field is in a sand that dips west and that is covered at an ancient erosion surface by nearly horizontal impermeable beds that have sealed in the oil and prevent its escape.

Briefly the conditions that are necessary for the formation of a large accumulation of oil are:

1. Strata containing organic matter to provide sources of oil.
2. A rock with openings to serve as a reservoir.
3. An impermeable cover to prevent the escape of the oil.
4. A suitable elevated structural feature to facilitate the separation of oil, water, and gas.[1]

Natural Gas.—Natural gas, or rock gas, is an aeriform mixture that is found at or beneath the surface of the earth and is used as an illuminant, for fuel, for generating power, and for making carbon black (soot). Most natural gas has a high heating value, and it is superior to artificial gas made from coal. Nearly every oil field yields some gas, although there are a few gas fields that produce no oil. As a rule the gas stored

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[1] The deepest well in the world is near Wasco, California. It is 15,004 feet deep and was drilled with rotary equipment. It got oil at 13,000 feet, approximately 2,500 barrels a day (May 1, 1938).

in the rocks is under strong pressure. When the natural reservoir is penetrated by a well, the gas issues violently, and at places it has caused great destruction. Because it is under high pressure, the gas well can be connected directly with a pipe line, and its own pressure serves as the means of transporting it. A large part of the natural gas that is produced contains a vapor of gasoline, and by compressing or freezing the gas or by passing it through a heavy oil this gasoline may be recovered.

In addition to the combustible gases, others are encountered in the earth. Among these are hydrogen sulphide, sulphur dioxide, and carbon dioxide. The latter is a heavy gas and therefore collects in ravines and low places in the topography where occasionally the quantity becomes sufficient to cause death to animals. Some of the petroleum wells of Colorado and Utah produce, with oil, large quantities of carbon dioxide at such low temperature that the oil comes out as a partly frozen slush that is very difficult to handle. In Utah, carbon dioxide is compressed to make dry ice. Certain wells in the United States produce helium. It is a light gas, is not inflammable, and is used for inflating balloons.

NON-METALLIC MATERIALS, EXCEPT FUELS

Non-metallic mineral products, excluding mineral fuels, include building materials, clay products, cement, lime, gypsum, salt, sulphur, mineral fertilizers, etc. The United States produces about a billion dollars worth of these materials annually, which is approximately the value of the metals produced.

For building stones nearly all consolidated rocks are used: limestone, marble, sandstone, granite, and other igneous rocks. Shales and poorly consolidated sandstone and certain igneous rocks are not good building stones, because they disintegrate readily. Shales metamorphosed to slates are used for roofing, and sandstones well cemented are good building stones.

Clay products include building brick, tile, pottery, porcelain ware, refractory brick, glass pots, etc. In general the molded clay is fired until the surface is fused so that the product will retain its shape. Both residual and transported clays are used for making brick.

Portland cement is a material made of limestone and clay or shale. These materials in proper proportion are mixed and burned to a clinker, and the latter is ground to powder. The moistened product sets, on drying, to form a strong rock-like substance. Lime is made by heating limestone, driving off the carbon dioxide. It is used for making mortar and in the chemical industry. Gypsum heated to drive off part of the water forms plaster of Paris, which, moistened, dries to form a solid white material. This material is too weak to make rugged structures but is widely used for interior finish of buildings. Sulphur and sulphides are mined for making sulphuric acid and other products. Common salt is obtained by evaporating salt water and by mining rock-salt beds (page 275) underground. Potash, bromine, and other salts are associated with rock salts and brines (page 275). Mineral fertilizers (page 53) include phosphates (page 275), nitrates, gypsum, and limestone.

Diamonds, emery, garnet, sandstone, and pumice are natural abrasives widely used. Other useful mineral substances include asbestos, fluorite, barite, mineral paints, fuller's earth, graphite, mica, quartz, feldspar, glass sand, molding sand, bentonite, lithographic limestone, lithium minerals, magnesite, gems, and many others.

ORE DEPOSITS

Ore deposits are parts of the lithosphere that may be worked commercially for one or more metals. They are formed by the same processes that form other rocks—that is, by gradation and vulcanism—like other rocks they are deformed by diastrophism, and where exposed at the surface of the earth they are weathered. Certain mineral deposits are igneous rocks in the strict sense. Other deposits are sedimentary beds that contain valuable materials. These include iron-ore beds, gold-bearing gravels, etc. In ore deposits the same materials are found that are present in other rocks, but they are sufficiently concentrated to be of economic value.

There is, however, an important group of mineral deposits that are neither igneous rocks nor sedimentary rocks. These are mineral veins which have been formed by water solutions moving through openings, chiefly through fissures, and depositing valuable minerals in and along them. These mineral-depositing waters are in part ordinary ground waters that contain the metals or other valuable materials in solution. The great majority of mineral veins, however, are found in and near intrusive igneous rocks, and their relations to igneous rocks are such as to warrant the conclusion that the veins are connected in origin with the intrusives. It is the belief of many students of ore deposits that solutions that deposited most of the veins have been derived from the cooling igneous intrusives.

Kinds of Ore Deposits

Magmatic segregations	Formed by consolidation of magmas
Pegmatites	Formed by "aqueo-igneous" solutions derived from igneous intrusives
Contact-metamorphic deposits	Formed by solutions from igneous intrusives replacing invaded rocks
Veins and similar deposits	Formed by mineral-bearing waters moving along fissures and other openings
Sedimentary beds	Formed by processes of aggradation

COMMON ORE MINERALS

Metal	Mineral	Elements present	Percentage of metal	Formula
Iron	Hematite	Iron, oxygen	70	Fe_2O_3
	Magnetite	Iron, oxygen	72.3	Fe_3O_4
	Limonite	Iron, oxygen, water	59 8	$2Fe_2O_3.3H_2O$
	Siderite	Iron, carbon, oxygen	48 3	$FeCO_3$
	Pyrite	Iron, sulphur	46 6	FeS_2
	Pyrrhotite	Iron, sulphur	60 4	Fe_7S_8
Copper	Native copper	Copper	100.0	Cu
	Chalcopyrite	Copper, iron, sulphur	34.6	$CuFeS_2$
	Chalcocite	Copper, sulphur	79 8	Cu_2S
	Cuprite	Copper, oxygen	88.8	Cu_2O
	Malachite	Copper, carbon, oxygen, water	57.4	$Cu_2(OH)_2CO_3$
Zinc	Sphalerite	Zinc, sulphur	67.	ZnS
	Smithsonite	Zinc, carbon, oxygen	52.	$ZnCO_3$
	Calamine	Zinc, silica	54.2	$Zn_2H_2SiO_5$
Lead	Galena	Lead, sulphur	86.6	PbS
	Cerussite	Lead, carbon, oxygen	77.5	$PbCO_3$
	Anglesite	Lead, sulphur, oxygen	68.3	$PbSO_4$
Tin	Cassiterite	Tin, oxygen	78.6	SnO_2
Silver	Native silver	Silver	100.	Ag
	Argentite	Silver, sulphur	87.1	Ag_2S
	Cerargyrite	Silver, chlorine	75.3	$AgCl$
Gold	Native gold	Gold	50–100	Au

COMMON GANGUE MINERALS

Mineral	Elements present	Composition
Quartz	Silica, oxygen	SiO_2
Calcite	Calcium, carbon, oxygen	$CaCO_3$
Dolomite	Magnesium, calcium, carbon, oxygen	$MgCO_3.CaCO_3$
Barite	Barium, sulphur, oxygen	$BaSO_4$
Fluorite	Calcium, fluorine	CaF_2
Feldspar	Potassium, aluminum, silicon, oxygen	$K_2O.Al_2O_3 6SiO_2$
	Sodium, aluminum, silicon, oxygen	$Na_2O.Al_2O_3.6SiO_2$
Garnet	Calcium, iron, silicon, oxygen	$Ca_3Fe_2(SiO_4)_3$
	Magnesium, iron, silicon, oxygen, etc.	Many of complicated formulas
Tourmaline	Iron, silicon, aluminum, boron, oxygen, etc.	Variable

An ore mineral is one that contains a valuable metal. In most deposits the ore minerals are associated with large amounts of material consisting of gangue and country rock. The gangue is the valueless material deposited along with the ore and usually is earthy or non-metallic. The country rock is the rock that incloses the ore deposit. In many deposits the ore grades into the country rock, and much of the latter is removed by mining the ore. Certain ore and gangue minerals are listed in a table on page 421. In some ores the valuable metal predominates, but in most deposits the ore minerals are present in subordinate amounts. An iron ore generally contains between 25 and 70 per cent iron, a zinc ore generally between 3 and 25 per cent zinc, and a lead ore between 3 and 15 per cent lead. Copper ores are generally low in copper, and the bulk of the metal is derived from ore that carries between 0.75 and 5 per cent copper. Silver ores carry between 8 and 30 ounces or more of silver per ton, and gold ores generally contain between $3 and $10 per ton gold. Many ores contain two or more metals. Thus the copper ores of Butte, Montana, carry important amounts of silver and gold, and there are few copper ores that do not contain small amounts of the precious metals. Lead and zinc are very commonly found in the same deposit, and lead and silver are common associates. In general, in the present practice of treating ores by milling and smelting them, two or more metals are recovered from the ore, and ores that contain less than the amounts of the metals stated above often are utilized.

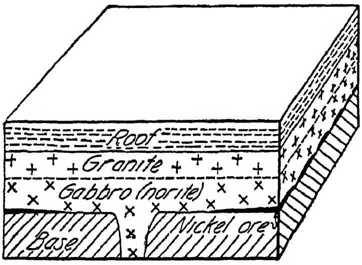

FIG. 451.—Sketch showing part of a laccolith after magmatic segregation. The lighter rock, granite, is at the top of the igneous mass, and below is gabbro with nickel ore at the base.

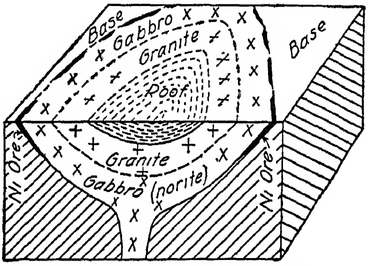

FIG. 452.—Sketch showing part of a laccolith after magmatic segregation, and down folding or slumping. The lighter granite is underlain and surrounded by the heavier, more basic gabbro; nickel ore occurs at the outer rim of the intrusive mass.

Magmatic Segregations.—Magmatic segregations are deposits formed by magmatic differentiation (page 320). The heavy differentiate generally is found below the lighter one (Fig. 451). Differentiation is most easily discovered and studied in laccoliths, because they have "floors" on which the heavier differentiates rest. The heaviest material sinks, the lightest material rises, and the constituents of intermediate weight occupy an intermediate position between them. At places where igneous magmas have risen to form laccoliths the latter have slumped down. Thus certain laccoliths are basin-like sheets, and after erosion they present at the surface a ring of the heaviest material, inside which is a ring of lighter material, and inside that a ring of still lighter material such as granite. The Sudbury, Ontario, eruptive mass is said to show such relations, as is illustrated by Fig. 319, where a heavy ore consisting of iron, nickel, and sulphur is found below gabbro, and above the gabbro is granite. Figure 452 shows a heavy nickel-bearing ore below a lighter norite (gabbro).[1]

In certain regions there are dikes of iron ore which fill fissures believed to extend downward to bodies of ore that have formed by the differentiation of a deeply buried and invisible rock magma. These dikes of ore, also, are believed to be magmatic differentiations. Deposits that have formed by magmatic differentiation include ores of nickel, iron, titanium, chromium, platinum, and subordinate deposits of copper and gold. Diamonds also and other gems are formed in magmatic segregations.

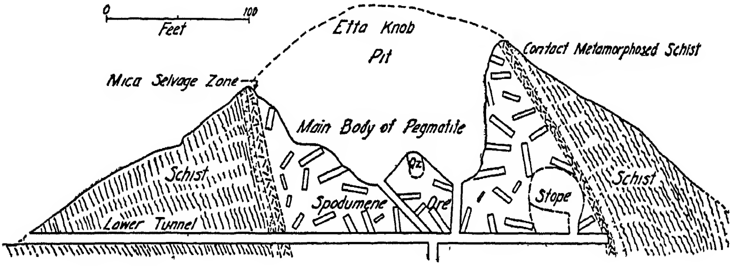

FIG. 453.—Geological cross section of Etta Knob pegmatite, southern Black Hills, South Dakota. The deposit is mined for spodumene which contains lithium. The large crystals are spodumene. (*After G. M. Schwartz.*)

Pegmatites.—Pegmatites (Fig. 453) are magmatic differentiation products, but they represent, in general, the lighter rather than the heavier products of the magma. Generally they are composed of large crystals, and of these feldspar, quartz, and mica greatly predominate. They have been called "giant granites." Some contain crystals of tourmaline (a boron mineral) and of apatite, which contains fluorine and chlorine. It is believed that the boron, fluorine, and chlorine, probably as gases, aided the growth of the large crystals, for it is known that these substances and also steam tend to keep the magma liquid, and that allows the crystals greater freedom of formation and results in the development of larger crystals. Because the parent magmas of pegmatites are believed to contain much water, they are called "aqueo-igneous" solutions. Pegmatites are found as dikes in the roofs as well as in the upper parts of batholiths. A few of them are associated with basic rocks. They are the chief sources of the micas, lithium minerals,

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[1] The largest nickel-ore deposits at Sudbury are in brecciated zones in older rocks below the gabbro. The relations are more complicated than is indicated by Figs. 319 and 452.

and feldspars of commerce; they contain also gems, such as tourmaline, ruby, diamonds, and other substances of value. On the other hand, pegmatites are rarely important as sources of the metals. A few are banded, like quartz veins, and some of them grade into quartz veins, but they are very rarely found grading into veins that carry commercial amounts of the precious metals.

Contact-metamorphic Deposits.—Contact-metamorphic deposits are replacements of invaded rocks formed by solutions that are expressed from the invading rocks. They are found in the garnet or other contact zones (page 387) in both sedimentary and igneous rocks but mainly in sedimentary rocks, particularly in limestones and calcareous shales (Fig. 423). Many of them lie against the intruding igneous rock, and they are rarely as much as one mile away from it. Contact-metamorphic deposits carry ores of copper, iron, zinc, and, more rarely, gold, silver, and lead. The gangue minerals include garnet, amphiboles, pyroxenes, and quartz. The sulphides, such as pyrite, chalcopyrite, and pyrrhotite; the oxides, such as hematite and magnetite; and the heavy silicates, such as garnet and amphibole, are mutually intergrown and have formed at about the same time. Some contact-metamorphic deposits follow beds, others cut across beds, but they do not follow well-defined fissures. In that they differ from fissure veins. The solutions that carried the ores, moved out from the intruding masses, and deposited material in the invaded rocks, were at high temperatures and under strong pressures, so that they seemed to be able to penetrate minute joints and the cleavage planes of minerals. For that reason the contact-metamorphic deposits are generally very irregular in shape and are rarely tabular, like veins. As a rule, the minerals of contact-metamorphic deposits form coarse aggregates.

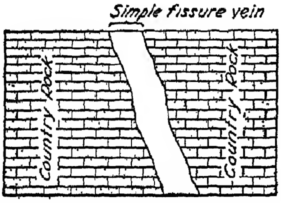

FIG. 454*A*.—Cross section showing a simple fissure.

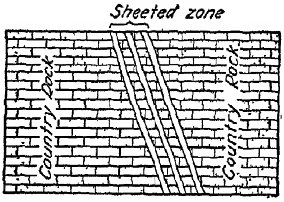

FIG. 454*B*.—Cross section showing a sheeted zone.

Mineral Veins.—Mineral veins are the most numerous mineral deposits and are among the most valuable ones. They are formed by waters moving through fissures and other openings, which deposit ore in the openings and soak into the wall rock, altering it and at places depositing ore in the wall rock near the openings by replacement. Veins exist in an almost infinite variety. They differ as to structure, texture, composition, and arrangement.

Structures of Veins.—The structures of veins depend largely upon the character of the openings that were prepared to receive the solutions that deposited the veins. Some veins fill single openings, as is illustrated by Fig. 454*A*. Others fill closely spaced parallel openings (Fig. 454*B*), which are sheeted zones. Still others fill irregularly fractured bodies of rocks and are called fractured zones. Along many veins the country rocks near the veins are partly replaced by ore. In certain irregularly fractured rocks small veins (veinlets) are closely spaced, and the rock is replaced between the fractures, so that the entire rock may be regarded as ore. Ore of this type is disseminated ore (Fig. 455). Some veins occupy fault fissures (Fig. 456); others follow certain beds that are brittle and easily fractured and therefore after movement offered favorable channels for waters. Certain beds are followed because they were easily replaced by the mineral-bearing waters that deposited the veins. Many deposits along fissures replace limestone beds below shales but do not replace the shales, because the shales are relatively impermeable (Fig. 457). Because the contacts of two rocks commonly are planes of weakness, they are often fractured and mineralized (Fig. 458).

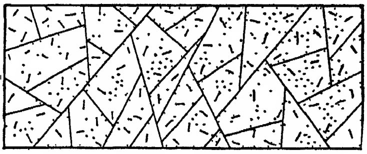

Fig. 455.—Figure illustrating disseminated ore. The rock is cut by many metal-bearing veinlets and, between these, contains dots and shots of metalliferous minerals.

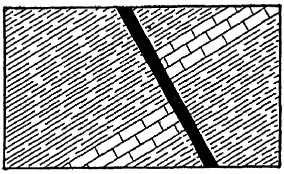

Fig. 456.—Cross section of a vein in a fault fissure.

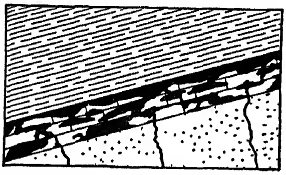

Fig. 457.—Cross section of a vein replacing a bed.

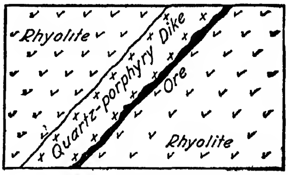

Fig. 458.—Diagram showing a vein at the contact between a rhyolite and a quartz-porphyry dike.

Textures of Veins.—Some veins are made up of one mineral or of two or more minerals so intergrown that the vein is essentially uniform throughout. Others consist of banded layers (Fig. 459). In certain veins these layers appear in the same order from the two walls to the center of the vein, that is, in symmetrical order. The waters moving through the fissure have deposited minerals layer upon layer on opposite sides of the channel, and the layers last deposited may come together and fill the channel. Where they do not come together, they leave an opening which is called a vug, or druse. Certain veins fill fissures without greatly altering the wall rock and without replacing it with ore. Such veins generally have sharp, regular contacts (Fig. 460).

Many veins do not have sharp, clean-cut walls, but they grade into the wall rock in such a way that it is difficult to determine where wall rock ends and vein begins. That is noted in many veins where the vein-forming waters soaked into the wall rock and replaced some of it by ore. The changes that are brought about in the wall rock by hot waters moving along fissures are due to hydrothermal metamorphism (page 387). A vein formed partly by replacement of the wall rock is a replacement vein (Fig. 461).

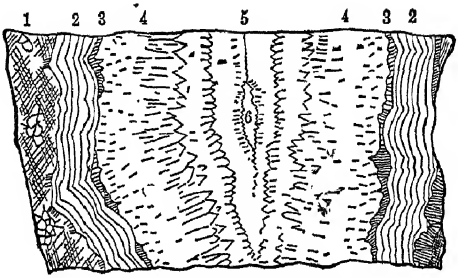

FIG. 459.—Section of a vein with symmetrical crustified banding. Creede, Colorado.

Composition of Veins.—Veins are the chief sources of most of the metals. The veins and nearly related deposits of ore formed in and along openings include the most valuable sources of gold, silver, copper, lead, zinc, mercury, and many other metals. Certain veins produce one metal; others produce two or more. In some veins the metals occur in the native state. That is true of most gold-bearing veins and of some copper-bearing ones. Many of the metals occur combined with other elements. Thus lead is found mainly as the sulphide, galena; zinc as the sulphide, sphalerite; and copper as the sulphides, chalcopyrite and chalcocite. Tin is found chiefly as the oxide cassiterite; and certain other

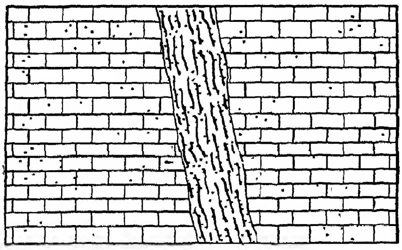

FIG. 460.—Fissure vein without replacement of the wall rock.

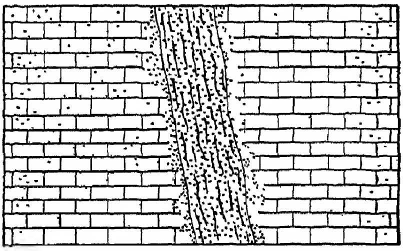

FIG. 461.—Fissure vein with replacement of wall rock (replacement vein).

metals often occur in compounds with arsenic. Iron is found in veins and beds. The iron-bearing minerals include the sulphides, pyrite and pyrrhotite, and the oxides, hematite and magnetite. In the great majority of veins the metals are present as sulphides or in association with sulphides.

Along with the compounds of the metals much valueless earthy material occurs. Such material constitutes the *gangue* of the veins. The most common gangue minerals are quartz and calcite, but veins contain also barite, fluorite, feldspar, and many other gangue minerals. As veins are worked out, often it is found that a single vein changes in composition at successive depths. Certain lead veins change downward to zinc veins; certain zinc veins become copper veins with depth; and certain copper veins become tin veins. The order is essentially the same if the same metals are present, and the reverse order is rare. A tin vein does not pass downward into a vein of copper, lead, or zinc, nor does a copper vein change to a lead vein with depth. These different parts of veins in which one metal predominates are called *zones*, and the zonal arrangement is a common feature of many veins. In some veins similar changes are found along the strike. Thus one may follow a vein along the surface through zones in which tin ore gives way to copper ore and copper to zinc ore. These arrangements of zones of ore are due to precipitation of different metals by solutions under changing conditions of temperature and pressure.

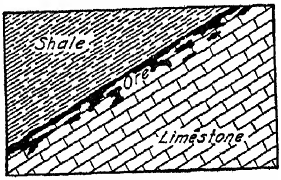

FIG. 462.—Ore deposit formed at and below the contact of shale and limestone.

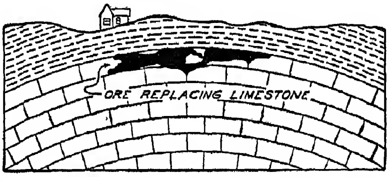

FIG. 463.—Cross section showing ore near the axis of an anticline replacing limestone below shale.

Sources of Vein-forming Waters.—Many veins are deposited by ordinary rain water, which soaks into the ground and dissolves metals from the rocks. The water circulating along cracks and fissures deposits the material it has gathered, and thus the ore is concentrated in and along the fissures. Veins of nickel ore in nickel-bearing basic rocks in New Caledonia have formed by this process, and iron and manganese ores have been concentrated in veins in rocks that contain iron and manganese. Certain valuable lead and zinc deposits generally are believed to have been concentrated by ground water and to have no connection with igneous processes.

The great ore veins are believed to have been formed, in the main, by ascending hot fluids that have escaped from cooling igneous masses. When deposits are associated with shale, they generally are found below the shale (Fig. 462). Shales are relatively impervious to water and form the great natural barriers to solutions. Limestones are replaced readily by ore; and if the ores are in limestone near shale, they are nearly everywhere below the shale, which suggests that the ore-bearing waters rose in the limestone and were halted by the shale where the ores were deposited. At places the rocks are arched, and the ores occur in anticlines in limestone below the shales. Ascending waters would converge in the upfolds and deposit ore below the impermeable shale barriers (Fig. 463).

The waters that deposited the ore veins are believed to have been hot, because the larger number of veins are associated with igneous rocks, and moreover the wall rocks near the veins show alterations that are characteristic of hot waters (hydrothermal metamorphism). Moreover, metallic sulphides are being deposited by hot waters today at Steamboat Springs, Nevada, and at other places.

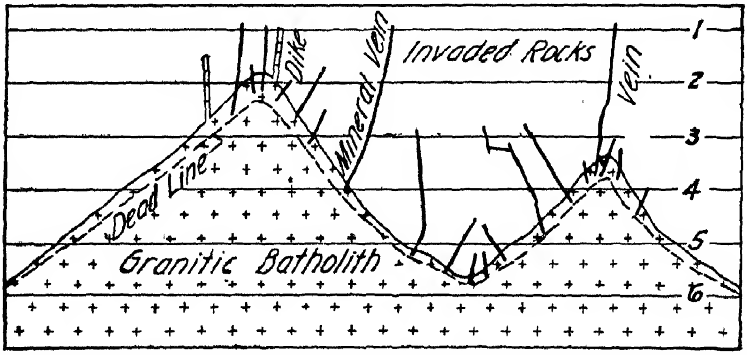

Fig. 464.—Cross section of a granitic batholith with mineral veins at its outer margin and in its roof.

While it is obviously not possible to observe the deposition of ores around deep-seated intrusives or to examine the whole of such intrusives and their associated ores, one may study many different intrusives and associated ores at many different stages of erosion. The study of many of these deep-seated intrusives, or batholiths, shows that they have upper parts called "roofs" that are very irregular and generally broad. Their contacts with invaded rocks generally slope away from the mass so that they become broader downward (Fig. 464). Many metalliferous veins are found in and around granitic batholiths, and it is believed that the solutions that deposited them were expelled from the cooling magmas that solidified to form the batholiths.

Sedimentary Deposits.—When rocks weather and are eroded, their materials often are separated. Thus a granite which is composed of quartz, feldspar, and mica will break down and form quartz sand and clay, which go to make up sandstones and shales. The iron in dark mica may be carried by streams to the sea and be deposited along with iron from other sources to form sedimentary beds that contain iron (page 276). The sand that is liberated is valuable for making glass and for other purposes, the clay for making pottery, and the iron ore for making iron and steel. In the granite these materials were not sufficiently concentrated to be of commercial value. Thus running water, the lakes, seas, etc., which collectively constitute the hydrosphere, may be regarded as a giant concentrating mill. By processes of gradation, deposits of coal, petroliferous beds, gypsum, iron ores, manganese ores, gold, platinum, tin, salt, phosphate rock, and many other valuable materials are formed. Some of these are carried along mechanically by the water and are deposited along streams and in the lakes and seas. Others, such as phosphate rock, and some iron ores are deposited by chemical precipitation from the waters. At many places gold, platinum, and tin ore are washed by water from the tops of deposits and are concentrated by water in ravines and creeks near their sources. At other places the materials that were eroded from the parent sources are carried far away. The concentration of materials in lakes and seas is treated on pages 214 and 275.

Deformation of Mineral Deposits.—Ore deposits, like other rocks, are deformed by earth movements. They are faulted or folded, or where deeply buried they may be deformed by dynamic metamorphism. Certain veins are broken into small blocks by systems of parallel faults; others are folded to form anticlines or synclines. All types of ore bodies may be deformed in various ways. When folded at great depths, ore deposits take on the characteristic structures of gneisses and slates. Thus certain iron ores of the Vermilion range, Minnesota, have been folded at great depth, and in the ore a slaty cleavage has developed by rearrangement of the flat crystals of hematite. Certain sulphide ores composed of quartz, hornblende, and pyrite have been deformed so that they now have the structure of gneiss

Weathering of Ores.—Ores, like other rocks, are subject to weathering at the earth's surface. Where exposed to the action of air and water, the deposits break down and form new minerals. In certain deposits, valueless materials are carried away by ground water, leaving the valuable material in a more highly concentrated state. In the Lake Superior iron-ore districts, beds of the iron-ore formations with about 25 per cent iron, by removal of material other than iron by surface waters, have been converted into ores with 50 per cent iron or more (Fig. 465). Near Little Rock, Arkansas, a highly aluminous igneous rock has been weathered and leached of silica so that it is now a high-grade aluminum ore.

Nearly all of the veins with ores of copper, silver, lead, zinc, and other metals contain sulphides. Pyrite, iron sulphide, is nearly always present. Where the veins are exposed at the surface of the earth, they are attacked by air and water, and they undergo a series of changes. Thus pyrite will be oxidized, forming sulphuric acid and iron sulphate.

$$FeS_2 + H_2O + 7O = H_2SO_4 + FeSO_4$$

The sulphuric acid that is formed will dissolve copper, zinc, and certain other metals, which are carried downward. The iron sulphate will be further oxidized and will break down and form iron hydroxide, which is insoluble and will remain at the outcrop. Thus nearly all deposits that carry iron sulphides will be marked by iron hydroxide, which stains the croppings of the vein, giving it a rusty appearance. This altered iron-stained material is called an "iron hat," or "gossan." As a rule, the outcrops of copper sulphide veins carry very little copper. Gold, unlike copper, is not soluble in sulphuric acid, and gold-bearing veins are generally as rich in gold at the outcrop as at depth or even richer. Copper is carried downward as copper sulphate, for in the presence of air and water copper is highly soluble in sulphuric acid.

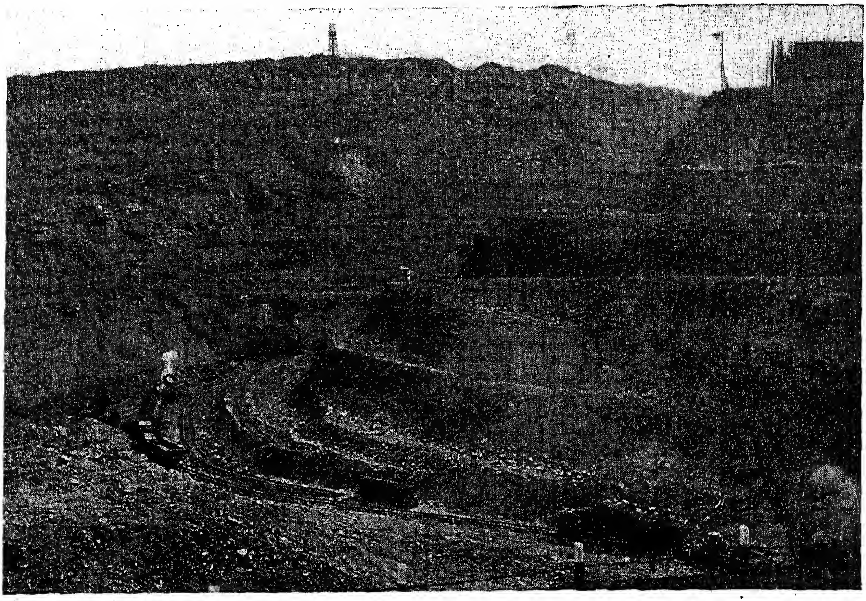

FIG. 465.—An open-pit iron mine on the Mesabi Range, Minnesota.

Where gold veins are exposed at the surface, the gold commonly will accumulate in the outcrop to be washed away by running water. Thus the particles of gold will be strung out along the surface below the vein cropping and will be washed into the beds of the streams, as is shown by Fig. 466. The stream gravels are washed into long wooden boxes, or sluices, and the gold, being heavier than rock, settles to the bottom of the box and is recovered. This method of recovering gold is "placer" mining, and gold-bearing gravels are placer deposits. All ore minerals that are heavy and not easily dissolved by ground water are likely to be accumulated in placer deposits. These include gold, platinum, tin oxide, diamonds, rubies, and other gems. Figure 467 illustrates the accumulation of a diamond placer by erosion of diamond-bearing material from a peridotite intrusive.

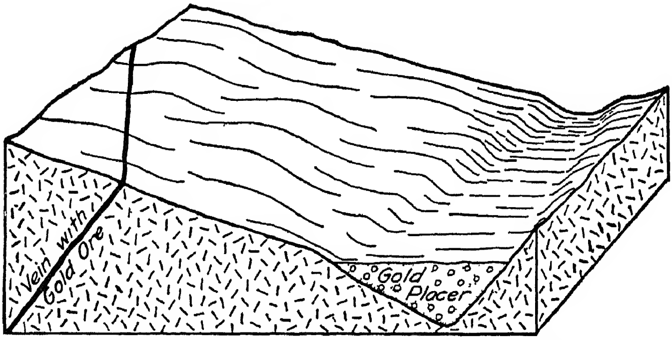

Fig. 466.—A vein with gold ore cutting granite. Due to degradation of the vein a gold-bearing placer deposit was formed along a nearby stream.

Sulphide Enrichment.—The water that dissolves copper and other soluble metals from the outcrops of copper sulphide deposits moves downward carrying the copper with it; and when it reaches the water level, it enters a changed environment. The water level, as already stated, is the level below which the openings in rocks are filled with water. The water seals out the air, which is present in openings above the water level, and consequently below the water level air is absent. Copper is readily dissolved by acid in the presence of air, but the solu-

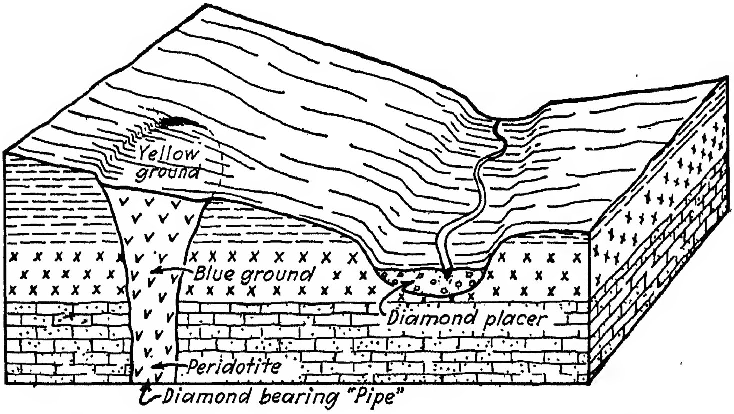

Fig. 467.—A diamond-bearing pipe. Diamonds removed from its eroded upper portion are deposited in gravels along a stream near the pipe.

tion in the absence of air loses its copper, which is precipitated by pyrite or by pyrrhotite or by many other minerals.[1]

As a rule, some copper carbonate or copper oxide will form just above the water level, so that the copper vein after oxidation and weathering will show a series of standard changes from the surface downward. This is illustrated by Fig. 468. Near the surface is found the gossan, or "iron hat," which carries little or no copper. A pit sunk through this will encounter first the oxidized copper ore with carbonates and oxides of copper. Still deeper it will encounter the level of ground water; and near that secondary copper sulphide ore will appear. At still greater depths the sulphide ore is found in its original state, and it is reasonable to suppose that the entire deposit from the surface downward was once like the primary ore and that it has been changed by ground water to the various types of ore that are found above it. The process by which the rich secondary sulphide ore is formed is called sulphide enrichment. Deposits of ores of other metals also show changes that are brought about by surface alteration, but each metal behaves in its peculiar way, depending upon its chemical properties.

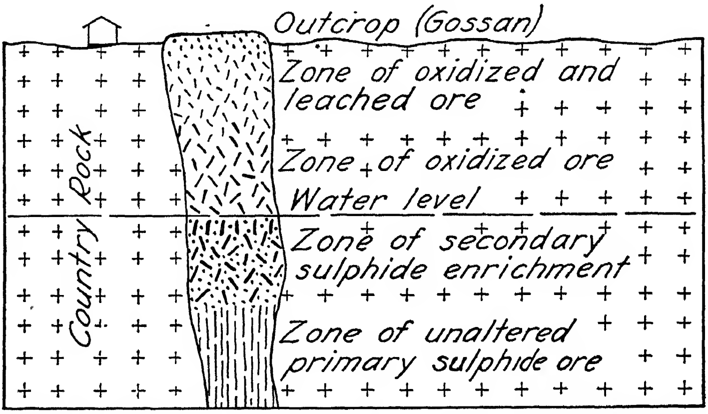

FIG. 468.—Diagram illustrating zones forming in a sulphide lode due to weathering and sulphide enrichment.

Summary.—There are certain ore deposits which have no direct connection with igneous activities. These include the sedimentary beds and the products of weathering and concentration of sedimentary beds and other rocks. Such deposits may be found wherever conditions were suitable for the deposition of such rocks. They commonly are situated in structural basins far from mountain ranges, and some of them are far from areas of igneous activity. Another group of deposits includes the veins and closely related deposits that have been concen-

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[1] Reactions forming copper sulphides are:

$$FeS + CuSO_4 \rightarrow CuS + FeSO_4$$
$$ZnS + CuSO_4 \rightarrow CuS + ZnSO_4$$

trated by ground water, which dissolves the metals from rocks in which they are scattered and concentrates them in and along the fractures through which ground water moves. These deposits often are found far from igneous rocks, and many of them seem to have no close relation to the distribution of igneous rocks.

The great majority of mineral deposits, however, are directly related to the igneous rocks and have formed by processes connected with igneous intrusions. These include the magmatic segregations, pegmatites, contact-metamorphic deposits, and the majority of mineral veins. Many veins are formed doubtless by solutions originating in igneous bodies. The intrusives contain water and other fluids which during cooling are expelled from the intrusives. Where they find suitable openings, they flow through them. On cooling or through reactions with the wall rocks the metals are deposited.

The mountain regions are, in general, the regions where intrusive rocks are most common and where strong fracturing and faulting of the rocks prevail. The intrusives provide sources of the metals, and the fractures afford openings for movements of fluids that deposit the metals. Consequently the mountain areas are generally the areas in which mineral deposits are most common, and these areas therefore supply the bulk of the metals. At certain places, however, the mountains have been worn down to near the elevation of the surrounding areas, and only their roots endure. At such places also mineral deposits often are concentrated, because the original deposits extended downward to depths below the present erosion surfaces.

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